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Some Plant-Soil-Water Relations in Watershed Management

LEON LASSEN, HOWARD W. LULL, and BERNARD FRANK,
Foresters

Division of Forest Influences
Forest Service

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PURPOSE

This publication has been prepared in response to the growing need by land managers, engineers, hydrologists, and others for more adequate information on how forest and range watersheds function, and how man's treatment of the land affects watershed functions. It attempts to bring together the available technical knowledge on the more important natural principles which appear to govern the interrelations of plants, soil, and water. It also provides a means for appraising the effects of given land conditions, treatments, and uses on stream-flow behavior.

The explanations and illustrations should help provide useful guidance to those technicians and administrators who are concerned with protecting, restoring, improving, or managing watershed lands wherever these may be found.

INTRODUCTION

The intimate relations of land and water, and the effects of land treatment upon the nature and quality of stream flow have been recognized and to some extent appraised as far back as the days of ancient Greece. With the passage of time, and especially since the full-scale development of our modern industrial civilization, the steadily growing need for the control and conservation of our soil, water, and other related vital resources has increased man's concern over the effects of land use upon floods, water shortages, and their associated difficulties. Out of this concern have come the principles of modern land conservation, and more recently, the concept of watershed management, which has for a major objective the application of these principles to drainage basins.

This broad objective is generally acceptable, and few people disagree that such management is needed. However, few understand the complex plant-soil-water relationships, both physical and biological, that underlie it. This publication describes some of these relationships, and indicates their practical significance with special reference to the protection and management of forest, brush, and range areas in the interest of their optimum long-run service to present and future generations.

The "why's" of the more significant effects of plant cover and cultural practices upon soil and water will especially be stressed. The authors believe such an approach to be essential because, although past research has emphasized—and to good advantage—to what extent cover and culture affect soil and water behavior, it has neglected in great part the "why's." As a result, most research findings have applied only to the small plots or other limited areas on which the investigations were undertaken. This situation in turn has led to unnecessary duplication of investigative work in an effort to widen the areas of application. Also, some investigations, because they were founded on inadequate knowledge of the fundamental processes involved, have produced results which have appeared contradictory to those obtained from previous studies.

Practical watershed management will be achieved only to the extent that it rests upon a thorough understanding of the fundamental natural processes which govern the behavior of soil, plants, and water. Fortunately, progress in the allied fields of soils and plant physiology, among others, and in the field of plant-soil-water relations as a whole, has provided some insight into these processes.

The complex interrelations of plants, soil, and water as modified by human activity involve the simultaneous interaction of many factors and processes. For ease of presentation it has been necessary to separate these into artificial classifications of subject matter. Such a separation represents merely a convenient device for examining the differences as well as the similarities among the various factors and processes, so that the relative significance of each and their impacts upon each other may be appraised more adequately. Actually no such division exists in

nature. The various processes to be discussed, such as evaporation, transpiration, infiltration, percolation, or such factors as soil depth or root depth, amount of soil organic matter, cutting practices, all blend into each other much as the colors in a spectrum.

Recognition of this blending of the processes and factors governing plant-soil-water relations is vital to an understanding and application of the principles and practices of watershed management. It is necessary to keep this in mind to avoid the mistake of seeking or relying upon "single-factor" solutions to watershed problems, no matter how simple the problems or solutions may appear.

The following sections will consider the relation of watershed management to stream flow; the function of soil in water storage; the effect of evaporation and transpiration on available storage space in the soil; the movement of water into and through the soil; and finally the application of these fundamental relationships to the actual practice of watershed management.

This publication can cover only a few of the more important factors governing plant-soil-water relations. Many are omitted or given only cursory treatment. For instance, little consideration is given to soil biota, the influence of plant succession, or interception of precipitation. Also, because attention is centered on water relations, erosion as a process and the erosion-control aspects of watershed management do not receive direct consideration. Although these factors are not specifically discussed, they cannot and should not be overlooked in considering specific watershed-management problems.

The examples and references in this discussion apply largely to forest and range areas. This is for several reasons: throughout the Nation these lands cover the major part of our watersheds, receive most of the precipitation, and produce most of the runoff; some of the more significant findings on plant-soil-water relations have come out of forest and range research; also, the illustrative material from these sources is well known to the authors. Nevertheless, because of their fundamental nature, the plant-soil-water relations described here apply equally to agricultural land. Thus crop and pasture vegetation influences stream flow and soil-water movement and storage, in much the same manner as does forest and range vegetation.

THE MEANING OF WATERSHED MANAGEMENT

The objective of watershed management is to meet the problems of land and water use, not in terms of any one resource, but on the basis that all resources are interdependent and must therefore be considered together. The complete drainage basin or watershed lends itself admirably to this approach. As a natural unit it reflects the interactions of soil, geology, water, and vegetation by providing a common end product—runoff or stream flow—whereby the net effects of these interactions on that product can be measured and appraised.

Stream flow in its various aspects is a recognized criterion of watershed conditions, and also of the effectiveness of watershed management. It provides useful indicators of the net results of conservation practices, including total yearly amount of flow, seasonal regularity, frequencies and extremes of high and low discharges—such as major floods and minimal flows—and quality.

The technical soundness of planning and developing river basins as units is now generally accepted. Since watershed management rests on the principle that water flow is a product of the land, it is considered by many people a chief requisite of a type of river-basin development in which conservation measures will be applied in conjunction with, or better still, in advance of engineering works for flood control, hydroelectric power, navigation, or irrigation. This alliance of land conservation and engineering can, for example, prolong the life of many reservoirs by reducing sedimentation, provide higher quality water for municipal and industrial use, reduce harmful flood discharges, and provide many other benefits.

Land Use and Stream Flow

A basic premise of watershed management is that the amount and rate of stream flow express the natural and cultural characteristics and conditions of the watershed which produces it. Graphically, these characteristics are expressed in the stream-flow hydrograph which represents rates of flow past a stream-gaging station over a period of time (solid line, fig. 1). Four factors affect the volume and rate of runoff and consequently the shape of the hydrograph: Precipitation (kind, amount, intensity, and distribution); drainage basin characteristics (including size, shape of basin, length and steepness of slopes, and stream density); soil and its plant cover; and changes in soil and cover through land use. So far as we know now the soil and its cover is the only factor subject to modification by man. The effect of soil-cover changes can be visualized by studying the stream-flow hydrograph.

Figure 1 shows a composite hydrograph representing typical stream discharge resulting from a high-intensity storm which produced surface runoff. The composite hydrograph, represented by the solid line, is actually drawn by the pen of the stream-flow recorder. The broken lines in the figure are not actually recorded by the pen on a chart. Rather they are derived by established techniques of hydrograph analysis to represent the several sources of flow which combine to make up the total flow past the gage.

How forest and range use affects stream flow can be visualized by separating the hydrograph into its component parts and analyzing the relation between each part and the character of the soil as modified by the condition and use of its vegetal cover. To help show these relations, frequent reference will be made to figure 1 and the illustration of rainfall disposal shown in figure 2.

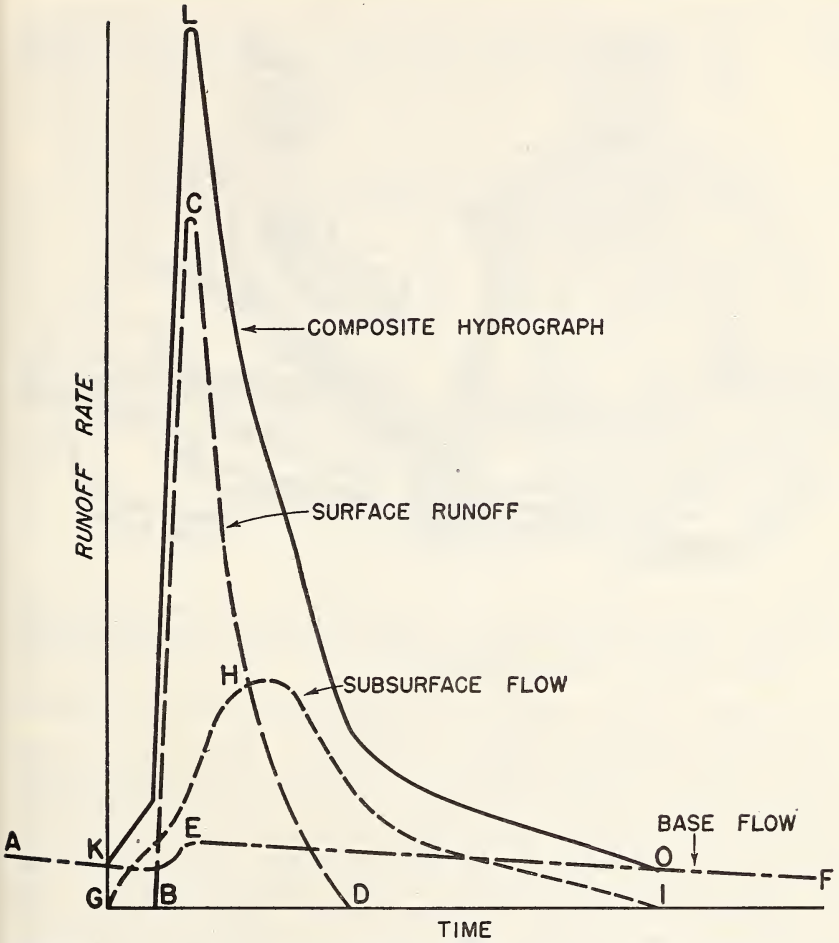

FIGURE 1.—A stream-flow hydrograph, with component parts, reflecting the effects of a rain.

The uppermost solid line in figure 1 shows total discharge past the stream-flow gage. This discharge comes from three sources: Base flow, surface runoff, and subsurface flow. Note that at any one instant the sum of base, subsurface, and surface flow equals total discharge, line K–L–O.

Base flow (line A–E–F in fig. 1, and fig. 2) comes from underground reservoirs which are the source of water for streams during rainless periods. This type of runoff responds much more slowly to precipitation than the other two sources. A second source is surface runoff, represented by line B–C–D and also shown in figure 2. This water runs over the surface of the soil. For this reason, surface runoff rates are very responsive to the intensity and amount of rainfall. The peak rate, as shown on the hydrograph (C), coincides in time and can closely approximate the peak rate of total runoff. The rate usually rises rapidly

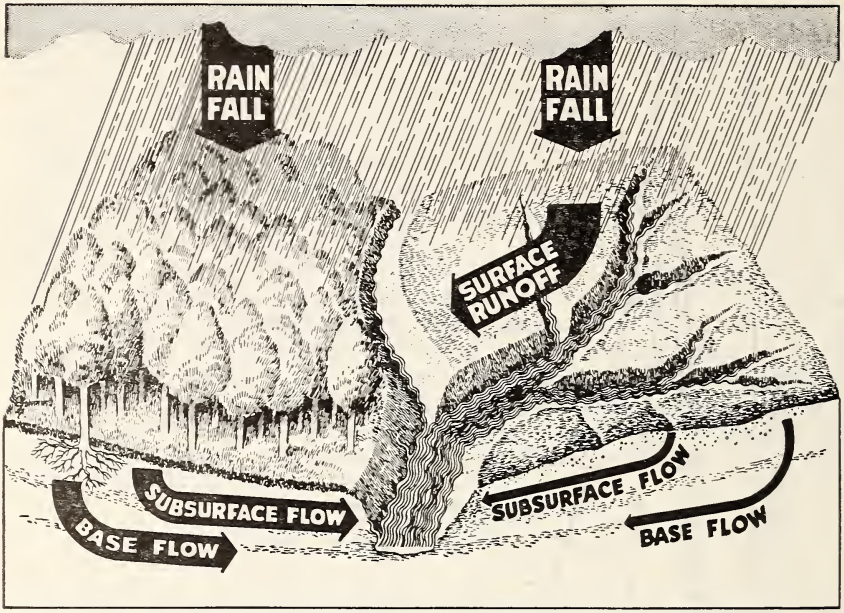

FIGURE 2.—Distribution of rainfall on vegetated and bare areas into surface runoff, subsurface flow, and base flow. Width of arrows indicates the relative amount of each component.

to the peak during or shortly after the intense part of a storm. Subsequent decline is somewhat less rapid.

Subsurface flow, the third component, is illustrated by line G–H–I, figure 1, and in figure 2. It originates from water stored temporarily in the soil at shallow depths (usually less than 24 inches) over a more or less impermeable layer. As water accumulates over this layer it will move down the slope through the soil pores towards the stream channel. Subsurface flow is not as sensitive to rainfall as surface runoff. Even for equal volumes of outflow its peak rate will be considerably lower than that of surface runoff, because the outflow is more prolonged.

The shape of the hydrograph of each component of stream flow reflects the time required for the water to reach the channel. Thus surface runoff, as shown in figure 2, takes the quickest way to the channels, often through existing or newly created rills and gullies on the surface. Subsurface flow moves more slowly on its way to the stream. Often, however, this type of runoff is intercepted by gullies before it reaches the channel and is then quickly transferred to surface flow. Base flow moves slowest. Once in the channel, the water from all sources moves downstream relatively fast. But the rate of movement of each type of runoff determines when it will reach the channel and consequently the shape the hydrograph will have.

Aside from the effects of vegetation on the interception of snow or rain, the amount of precipitation that goes into base, surface, and subsurface flow, respectively, is determined in large part by the physical characteristics of the soil. As will be

shown later, these physical characteristics are affected by the nature of the vegetation the soil supports. For this reason, any practices which change the nature of the vegetation will in turn cause changes in the physical characteristics of the soil and therefore in stream flow.

The physical characteristics of the soil determine its hydrologic characteristics; namely, infiltration—the rate at which water moves into the soil, percolation—the rate at which water moves downward through the soil, and storage capacity. These factors in turn determine the proportions of the three stream-flow components. When infiltration and percolation rates and soil-storage capacities are high, immediate runoff from a given storm will be low because the amount of surface runoff will be small. Under these conditions, the storm precipitation will move into the soil. There part of it may be stored, a part may go to subsurface flow, and still another may move on downward to base flow. These are the slow routes to the stream channel. Conversely, at the opposite extreme of low infiltration rates with little storage, surface runoff will account for much more of the precipitation, with a consequent marked increase in peak rates of stream flow.

An example of the effects of land use on stream flow is shown in figure 3.[1] The hydrographs from two of the several experimental watersheds at the Coweeta Hydrologic Laboratory in western North Carolina illustrate the changes brought about by mountain farming. The treated and the control watersheds cover 23 and 39 acres, respectively.

Stream flow on the treated watershed was measured for 6 years prior to its treatment in order to determine its runoff characteristics. The timber was clear-cut in 1939-40. Six acres were planted to corn and cultivated according to local practices. Twelve of the remaining 17 acres were grazed. Five acres were found too rough for agricultural use. The effects on stream flow are shown by a comparison of two storms of about the same size, one occurring before treatment in 1934, the other after treatment in 1946. The hydrograph of the control watershed shows little difference in stream flow from the two storms. Peak discharges were 12 and 15 cubic feet per second per square mile (c.s.m.).

Hydrographs from the treated watershed tell quite a different story. Whereas the 1934 storm produced a peak discharge of 16 c.s.m., the 1946 storm produced about 69 c.s.m. The difference was due almost entirely to an increase in surface runoff as shown by analysis of the hydrograph.

The above example applies only to the effects of a given storm on rates of stream flow. The effects of watershed treatment on the amount of water yield are not taken into account here. As will be shown, watershed management may affect this factor also.

As previously indicated, the causes of variation in stream flow resulting from land treatment must be sought in the funda-

[1] Unpublished data, Coweeta Hydrologic Laboratory. 1950. U. S. Forest Serv., Southeast. Forest Expt. Sta.

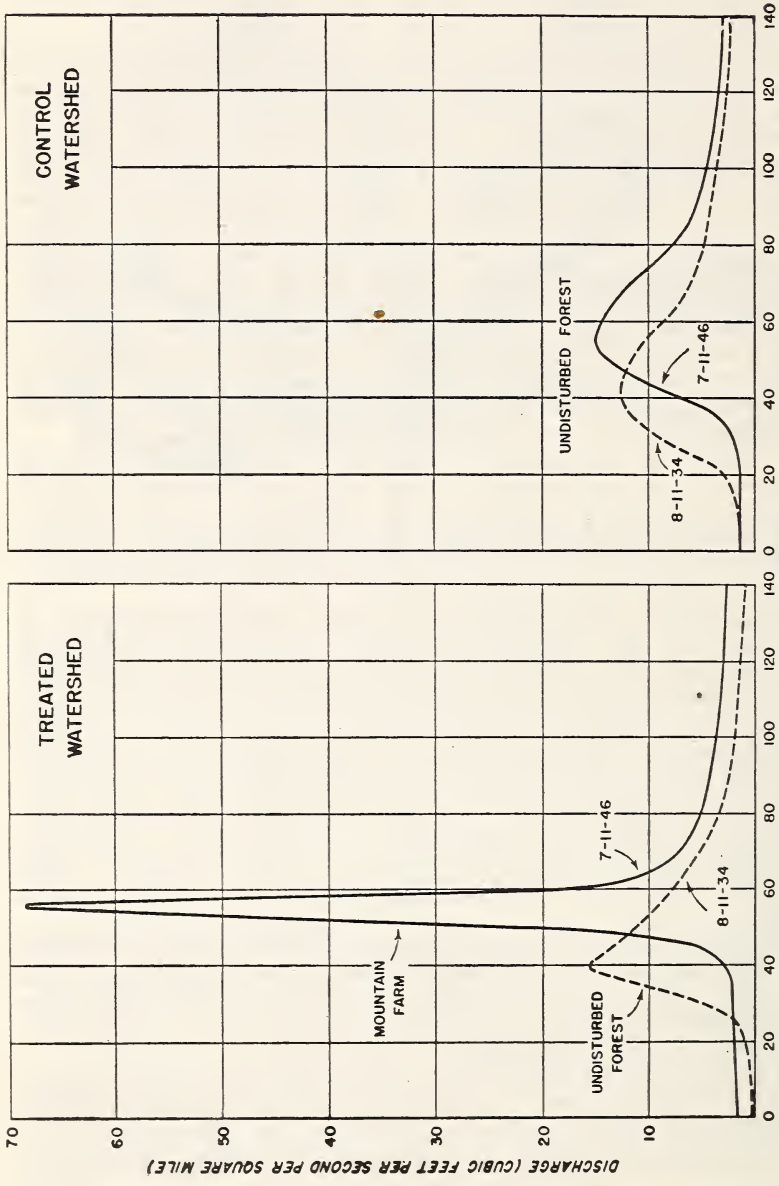

FIGURE 3.—Comparative effects of mountain farming and an undisturbed forest on stream flow.

mental properties and interreactions of plants, soil, and water. In this connection, the function of soil as a storage reservoir will be first considered.

THE SOIL AS A STORAGE RESERVOIR

For watershed management purposes soil may be defined as that portion of the weathered rock material which is subject to seasonal changes in moisture and is occupied, or capable of being occupied by roots. The effects of land use on the hydrologic characteristics of the soil are most pronounced in this zone. It consists of a porous framework made up of mineral particles and organic matter, and pore space which is occupied by air and water.

Soils are not static. They are dynamic in the sense that they are continuously changing in color, depth, and structure in response to the activities of living organisms: Roots, burrowing animals, micro-flora and micro-fauna. They swell with the addition of water; they may shrink and crack as they dry. Frost expands them. They creep downward on steep slopes. Even the most stable soils very slowly move downhill. More rapid movement occurs with accelerated erosion, and the most rapid with running water or wind.

Mineral soil material is developed from the weathering of rock by physical and chemical processes at rates governed to a large extent by the climatic and biotic environment. These processes of soil formation are continually at work and, unless major changes in the environment occur, weathering of the underlying rock (parent material) gradually deepens the soil mass. This is a very slow process. Accompanying this process are other processes, such as geological erosion and leaching, which preclude establishment of a static state in the material affected by weathering. Effectiveness of all these various processes decreases with depth to the vanishing point.

Not all soils are formed in place. By far the greater part of the land surface is underlain by loose sediments or rock waste that have been transported by wind, water, or simply by gravity. The mechanical separates of which these transported and redeposited materials consist are, of course, the products of weathering and still are subject to further alteration in their new place of rest.

Wherever a soil supports vegetation, it is enriched to some degree by organic matter. Plant roots mix intimately with the mineral particles, die, decay, and become incorporated in the soil. Above ground the periodic fall of dead leaves, the decaying wood of dead branches and fallen trees, together with animal remains, add organic matter which in time also becomes part of the soil. These additions increase the soil volume, deepening it more rapidly than does the weathering process. An extreme example is the formation of pure organic layers several inches deep directly on bare rock. These function hydrologically much as mineral soils do. As will be shown later, organic matter has

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many significant effects on water movement and storage in the soil.

The effects of land use and vegetation on the soil are reflected principally in changes in the structure, number, size, shape, and stability of the aggregates. These changes in turn affect the size, shape, and distribution of the soil pores. These pores provide space for the storage of water and serve also as avenues for water movement. Varying greatly in size and shape, they compose roughly about one-half of the soil volume. The mineral and organic particles comprising the other half in effect serve as a framework for the pores. Differences in size and shape of the pores determine the amount of water which can be held in them and the rate of movement through them.

Hydrologic Significance of Soil Composition

The arrangement of the mineral framework, organic matter, and pore space varies considerably within and between soils. Within the root zone—as observed on the side of a freshly cut trench or roadbank—soil usually develops a certain profile characterized by several layers or horizons that differ from one another in color, texture, structure, and consistency. The individual properties of these layers are acquired as a result of the differential alteration of the "original" materials by weathering. Furthermore, the "original" materials themselves may consist of several layers of sediments of different characteristics, as in the case of stratified transported deposits. These various layers, whether they are developed by differential weathering in place or result from stratification of parent material, concern hydrologists because they affect water movement and storage.

Soil depth is an extremely important hydrologic characteristic because it affects, among other things, the storage capacity of the soil. All other things being equal, a soil 6 feet deep can hold twice as much water as a soil 3 feet deep. Of course, soil-storage capacity depends on many other factors besides depth, for example, texture, structure, organic-matter content, root depth, among others. Nevertheless, depth must be recognized as an important and often limiting factor in soil-water storage.

In respect to storage capacity, the hydrologic depth of a soil may be entirely different from its linear depth. Hydrologically speaking, a "shallow" soil may be regarded as one with limited storage capacity even though it may be several feet deep. A shallow soil can be considered as having less than 3.0 inches retention storage capacity. Such a soil may vary in depth from approximately 1 foot for a clay loam to 6 feet for a fine sand.

The uppermost layer of the soil under an undisturbed hardwood forest consists of pure organic matter (largely undecomposed leaves and dead branches), generally varying in depth from 1 to about 6 inches. Underlying this is a dark layer, one to several inches deep, consisting of a mixture of decomposed organic matter and mineral particles. The organic content may be as much as 20 percent by weight and very nearly 100 percent

by volume. Such soils are highly permeable, as a rule. Beneath these organic layers are one or more layers of mineral particles from several inches to several feet deep, more or less permeable to roots and water and with an organic-matter content rapidly decreasing with depth.

By contrast, cultivated soils commonly lack surface layers of organic matter. The upper layer to the root depth may contain small amounts of organic matter, perhaps from 3 to 10 percent by weight in the upper 3 to 4 inches, to less than 3 percent at lower depths. Some cultivated soils may develop so tight a layer at plow depth as to discourage both root penetration and water movement.

Living and "dead" organic matter in the soil is highly complex. In size it ranges from microscopic bacteria to burrowing woodchucks. Roots have great influence. They can exert a pressure up to 75 pounds per square inch, enough to raise sidewalks or crack rocks (*29*).[2] They pierce the soil, breaking up clods. They take up a considerable amount of room. Common evidence of their bulk is the slight elevation of the root collar of trees above the surrounding surface.

Cavities left by dead roots significantly affect water movement. They are highly permeable, since they are generally filled with a mixture of organic matter and mineral particles that have worked in from the surface layers. In relatively impermeable soils they provide the principal avenues of water movement and temporary (detention) storage.

Soil as Both a Sieve and Reservoir

Hydrologically, soil is an agent which holds water or passes it on to stream flow. In these capacities it can be likened to a series of sieves. When water is poured on an ordinary household sieve it readily goes through, only a small portion remaining on the meshes. When several sieves are nested together and water is poured through, it still goes through rapidly but somewhat less so than in the previous example. Also, more water is retained on the margins of the meshes. As the operation is repeated with sieves of different size meshes it will be found that the greater the mesh size the faster the water will go through and the less will remain on the mesh.

The series of nested sieves may be said very roughly to represent the soil. The greater the number of sieves the deeper the soil, and the greater the mesh (pore) size the more porous is the soil. When water is poured through, the sieve with the smallest mesh (pores) will control the rate at which the water passes through—just as the least permeable soil layer ultimately sets the rate at which water moves through the several horizons.

The foregoing analogy illustrates two processes which occur in the soil: Water storage (the amount adhering to the sides of the mesh), and water movement. However, it brings out only one aspect of storage, namely, capillary or "retention" storage

[2] Italic numbers in parentheses refer to Literature Cited, p. 62.

(water retained in the soil). This term refers to the water which is held by the soil particle against the force of gravity. This water also moves, but so slowly as to be considered for all practical purposes as being in storage. Water in retention storage is available for use by vegetation and for evaporation. It is unavailable for stream flow.

The noncapillary or large pores of the soil provide another form of storage. Here water moves downward under the force of gravity through pores whose diameter is greater than 0.05 mm. Because this water is only temporarily detained, it is commonly referred to as "detention" storage. This is the water available to stream flow.

The difference between detention and retention storage may be illustrated by a second analogy, namely, storage in an artificial reservoir. When a reservoir is full so that water stands level with its outlet, it may be considered to have its retention storage capacity satisfied. During a heavy rain additional water is stored in the reservoir, but only temporarily, that is, for whatever time it takes the water level to recede to the height of the spillway. During this period the additional water has been held in detention storage. Soils operate in a similar manner. The small pores provide retention storage capacity whereas the larger pores provide only temporary storage, merely detaining the water until it flows out of the soil.

Water retained in storage in the soil reservoir is removed by evaporation and by transpiration (extraction of moisture from the soil by plant roots and its later evaporation through leaves and other plant parts). By affecting the amount of water stored in the soil at a given time these processes consequently affect the amount of storage space available (the storage opportunity). Because transpiration is the more important of the two, removing water more rapidly and generally to greater depths than evaporation, the amount of storage available at any given time is affected by changes in transpiration due to the effect of land use on vegetation. Further, as will be shown, vegetation also affects the detention storage capacity of the soil as well as the rate at which water moves through it. These rather complex interrelationships of plants, soil, and water can best be discussed by considering one relation at a time and the forces involved. Retention storage will be given first attention.

Retention Storage—How Water is Held in the Soil

Water is retained in the soil by the adhesion of moisture to the surface of soil particles. This property of practically all solids to adsorb soil moisture on their surfaces is well known and has been put to many uses. The "humidity stick" used for estimating forest fire weather danger is one example. Another is the use of silica gel to dry out damp atmospheres of cellars and closets, and to maintain dryness in packaged machine parts.

Molecular attraction between like and unlike molecules holds water to the surface of a soil particle. The force with which water

is held decreases with the distance of the water molecule from the surface of the particle. In figure 4, *A*, the black disk represents a soil particle; the other circles represent successive films of moisture around the particle. The innermost film (*a*) is held tightly to the particle. It would require a rather large force to separate the moisture from the particle. The second film (*b*) is not held so tightly and the third (*c*) still less tightly. Because these films are held to the particle against the force of gravity, the holding forces must be greater than the force of gravity. At some point, however, the film becomes thick enough so that the holding force just equals the gravitational force. The amount of water adsorbed at this point is called the "field capacity" of the soil.

Since adsorption of moisture is a surface phenomenon, the total adsorbed on any particle at any adsorptive force is directly proportional to the surface area of the particle. Thus the surface area of the soil particles becomes an important factor in storage. Another important and related factor is that the surface area of a unit volume of soil particles will increase as the number of

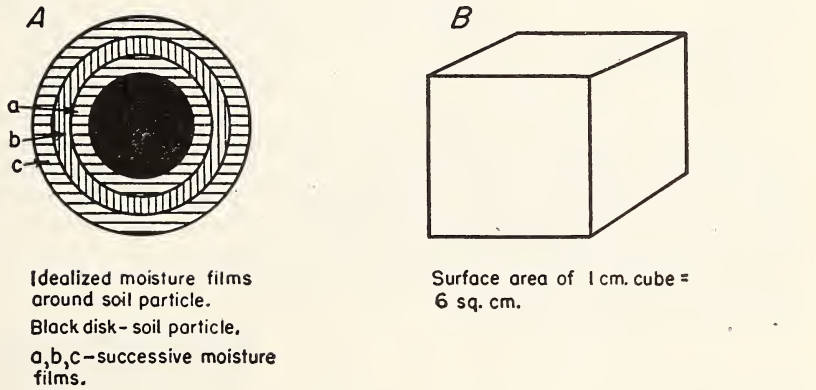
A
a
b
c
Idealized moisture films around soil particle.
Black disk - soil particle.
a, b, c - successive moisture films.
B
Surface area of 1 cm. cube = 6 sq. cm.
C
Surface area of 1 cm. cube split in two = 8 square cm.
D
Surface area of cube is increased by irregularities.

FIGURE 4.—Water films on a soil particle, and the relation of particle size and shape to surface area.

particles increases. For example, figure 4, *B*, shows that a 1-centimeter cube has a total surface area of 6 square centimeters; if the cube is split into 2 pieces (fig. 4, *C*) the surface area increases to 8 square centimeters. If the subdivision is continued so that cubes of 0.0001 centimeter dimension are formed (about the size of fine clay particles) the total surface area would be 60,000 square centimeters (*5*). The increase in area, by providing a greater space for adsorption, increases the amount of water which can be adsorbed. This applies to any value of the adsorptive force; that is, with increase of surface area the volume of water stored on the soil particles will be increased for all ranges of adsorptive force up to the point at which field capacity is reached.

Since the number and size distribution of soil particles in a unit volume determines soil texture, texture determines the amount of surface area and, in turn, water-holding capacity. Differences in water-holding capacity are indicated in the following tabulation of the number and surface areas of sand, silt, and clay particles in a cubic inch of soil with 50 percent pore space.

Texture:	*Diameter of sphere (mm.)*	*Number of particles*	*Surface area (sq. ft.)*
Coarse sand	1.0	1.56×10^4	0.53
Very fine sand	.1	1.56×10^7	5.29
Silt	.01	1.56×10^{10}	53.05
Clay	.001	1.56×10^{13}	530.49

Soil particles have thus far been assumed to have smooth surfaces. Under actual conditions the surfaces are indented; hence the adsorbing area is much greater. For example, if the cube in figure 4, *B*, had surface irregularities as shown in figure 4, *D*, its surface area would be considerably greater.

The retention storage capacities of several textural classes of soils are as follows (*2*):

Textural class:	*Retention storage capacity (inches depth of water per foot depth of soil)*
Fine sand	0.5
Sandy loam	1.7
Silt loam	2.5
Loam	3.3
Clay	4.5

Clay can hold 9 times as much water as can fine sand against the force of gravity. As the textural classes approach clay, differences become less.

How Organic Matter Affects Retention Storage

Retention storage is affected not only by texture but by the amount of organic matter as well. The effect of organic matter is to increase the storage capacity of the soil particle, and to increase the total volume of soil.

Organic matter has a high adsorptive capacity. In the colloidal

form it takes up as much as 4.4 times its own weight of water (*27*). When decomposed and mixed in the soil it coats the mineral particles with a gel-like, porous and highly adsorptive substance. Clinging to the particle, this material, in effect, increases its surface area and therefore its storage capacity.

The incorporation of decomposed organic matter into the soil may either increase or reduce retention storage capacity per unit volume of soil, depending on soil organic content. When small amounts of organic matter are added to the soil, the organic coating on the soil particles may not be sufficient to change their volume appreciably, so that the number of particles per unit volume remains the same while the surface area increases by reason of the organic coating. Under these conditions, retention storage capacity per unit volume would increase. Thus a study by the Northeastern Forest Experiment Station shows little relation between volume weight and organic matter content when the organic content is in the neighborhood of 5 percent or less.[3] This indicates that the number of soil particles per unit volume has not changed. Actually, however, the increment of organic matter should increase the volume weight, although the weight involved is probably too slight to give a clear-cut relation between volume-weight and organic content.

The same study also shows that as organic content increases above 5 percent, volume weight steadily decreases, indicating fewer soil particles per unit volume. Under these conditions, undoubtedly, particle displacement occurs because the coating of organic matter increases the volume of the individual particles. With fewer particles per unit volume less surface is available for water adsorption. Consequently, even though the storage capacity of the individual particle is increased, the capacity per unit volume of soil may be reduced. Thus, compacted soils and soils with low organic-matter content, because they have more soil particles per unit volume, may often have a greater retention storage capacity per unit volume than soils of similar type but with a larger organic-matter content. For instance, Hoover (*23*) reports a retention storage capacity of 7.93 inches in the upper 24 inches of an undisturbed hardwood forest soil; under an old-field loblolly pine stand a similar soil had a retention storage capacity of 8.54 inches. The hardwood site had a mull type humus (incorporated into the mineral soil) with relatively high organic content; a mor type humus with little incorporation of organic matter into the soil had developed underneath the pine stand.

With addition and incorporation of organic matter, soil must expand. Because the opportunity for expansion is chiefly upward, the soil will be deepened. Accompanying this is an increase in the retention storage capacity for the depth affected by organic matter.

An example of increased soil depth resulting from additions of organic matter is provided by the following comparative data

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[3] Original data used in preparing Station Paper No. 39, EFFECT OF SOIL AND COVER CONDITIONS ON SOIL-WATER RELATIONSHIPS. 1951. U. S. Forest Serv. Northeast. Forest Expt. Sta.

for the soils of two adjacent forest tracts, grazed and ungrazed woodland.[4] All percentages are on a volume basis and refer to the humus depth.

	Grazed	*Ungrazed*
Humus depth	2 inches	4 inches
Organic content	5 percent	20 percent
Total pore space	51 percent	63 percent
Field capacity	37 percent	38 percent

The ungrazed woods had also been grazed at one time, but its soil had improved considerably after a period of protection. The question to be answered is, How much deeper is its soil by reason of the greater humus depth? Also, how much was the storage capacity increased by this increment to depth?

Since the humus layer consists of organic matter, space, and mineral particles, it is easy to determine the volume of the mineral soil with the other two items known. This volume amounts to $4-4\ (0.20+0.63)=0.68$ inch, the volume equivalent of the mineral particles distributed throughout the 4-inch humus layer. These particles were removed from the mineral soil.

Correspondingly, the 2-inch humus layer of the grazed soil contains a volume of mineral particles equivalent to 0.88 inch. At the rate of 0.88 inch of mineral particles to 2 inches of humus, the depth of humus on the protected area while it was being grazed was $0.68/0.88 \times 2$ or 1.55 inches. Since under protection humus depth is 4 inches, the increase in depth would be $4.0-1.55$ or 2.45 inches. The resulting increase in retention-storage space is then 2.45×0.38 or 0.93 inch.

Thus soil depth may be increased by the incorporation of organic matter in the top soil layer and by the addition of organic matter to the top of the soil as well as by weathering at the bottom, and in a much shorter time. Increase in soil depth by decomposition of parent material is a relatively slow process. Moreover, the storage capacity of given increments does not equal that of humus.

INFLUENCE OF VEGETATION ON RETENTION STORAGE OPPORTUNITY

As has been shown, vegetation affects the amount of water that can be stored in the soil through the influence of organic matter on the storage capacity of the soil particle and on soil depth. A second way in which vegetation contributes is by taking water out of storage by transpiration. At any one moment the opportunity for retention storage in the soil depends not only on its water-holding capacity but also on the amount of water already in storage. The amount of retention-storage opportunity at any one time is the difference between the field capacity of the soil and its moisture content. The influence of vegetation on soil-moisture content provides the basis for evaluating its effect on retention storage.

[4] See footnote 3, p. 15.

Water is removed from retention storage by evaporation and transpiration. Although the processes are often considered as one, the marked differences in the manner and the degree to which they affect soil-moisture content requires that they be considered separately.

Factors Affecting Evaporation

Removal of soil moisture by evaporation provides storage opportunity in the soil for precipitation. Evaporation represents the movement of moisture in the form of vapor to the atmosphere from the surface of soil particles. In air the water-vapor particles move from areas of high concentration to low concentration, i. e., from high to low vapor-pressure gradients.

Rates of evaporation depend upon temperature, wind movement, and vapor-pressure gradient. Since all three elements are affected by the type and density of vegetation, any cultural or natural factors which alter vegetal characteristics will also affect the rate of evaporation. Vegetation, by shading the soil, reduces its temperature, thus reducing evaporation. Vegetation, by slowing down wind movement, retards the replacement of moist air with drier air and thus impedes evaporation. These effects, plus that of the insulating effect of litter against air movement and high soil temperatures, tend to keep evaporation losses to a minimum under dense vegetal cover. Evaporation under forest cover is generally less than under a grass cover, reflecting principally the greater shade produced by the trees and the insulating effect of the ground cover.

Evaporation losses are most rapid from the surface because the particles there are most directly exposed to wind movement and high temperature. The rate of movement of the vapor particles from lower layers of the soil to the surface is determined by the size and shape of the pores. The greater the number of large pores the more direct the routes by which vapor can escape. Sands and other soils with numerous large pores or cracks are therefore subject to higher rates of evaporation than are fine-textured soils.

The depth to which evaporation from a bare soil extends depends on soil porosity and depth to water table. On fine-textured upland soils where the water table does not influence evaporation, water losses are limited generally to the first foot of soil. Coarse-textured soils and cracked soils possess more and larger avenues for escape of the vapor particles, and here evaporation may remove water from depths as great as 5 to 6 feet.

Where the water table lies close to the surface so that capillary flow feeds water to the surface for evaporation, soil porosity again is the limiting factor, but this time in relation to the movement of water in its liquid state rather than in its vapor state. Since small pores conduct capillary water farther than large pores, evaporation from fine-textured soils affects water tables to greater depths than in coarser-textured soils. For a coarse sand the limiting depth is about 14 inches; for a clay, 3 to 4 feet (*33*).

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Thus, in areas where evaporation opportunity is great because of high water tables, the depth to which evaporation is effective is greater in clay than in sand; whereas in upland soils, where water tables are not a factor, the reverse applies.

The amount of water evaporated varies not only with temperature and wind movement, but also with the supply of available water (evaporation opportunity). This factor outweighs all others. Even during the high temperatures of summer, evaporation losses will be no greater than during the cooler winter months if the soil is not frequently wetted by rainfall. During winter, when the soil is usually at or near field capacity, evaporation rates will be limited, not by the supply of water, but by conditions due to low temperatures.

The close dependence of evaporation upon the supply of available water is often confused with the relation of evaporation to temperature. This error originates in the method of determining evaporation by means of an evaporation pan with a free water surface. Under such conditions, water loss is closely related to temperature because a continued supply of water is available. If all the water were evaporated from the pan, obviously the relationship would no longer apply. Similarly, high soil temperatures cannot affect evaporation when no moisture is present. Thus, total evaporation losses are determined by the quantity of water that is available.

It is well established that evaporation rates vary considerably, according to the degree of protection afforded the surface by vegetation (*28*). It is safe to assume, however, that once a cover is established, little variation in evaporation rates is likely to occur under forest and range management practices designed to maintain sufficient vegetation to stabilize the soil at all times. Changes in water yield achieved by altering vegetation will result largely from the effects on transpiration rather than on evaporation.

Plants Pump Water out of Storage

Transpiration, like evaporation, provides storage space in soil. In this respect plants may be likened to pumps emptying a reservoir. The more water that is pumped out, the more space is available in the reservoir for precipitation. How fast and how much water can be pumped out depend in part on physical and biological limits, and in part on land-use practices. The physical limitations to transpiration will be discussed first.

The availability of water for plants is related to the force with which the moisture is held to the soil particle. It can be expressed in units of force required to separate the moisture from the particle. These units will be expressed here in atmospheres, although they are frequently expressed also as heights of equivalent unit columns of water, or sometimes as the logarithm of the column height (pF).

As shown in figure 5, the force required to separate moisture from soil increases rapidly with a decrease in moisture content below field capacity. At field capacity, the force is equivalent to

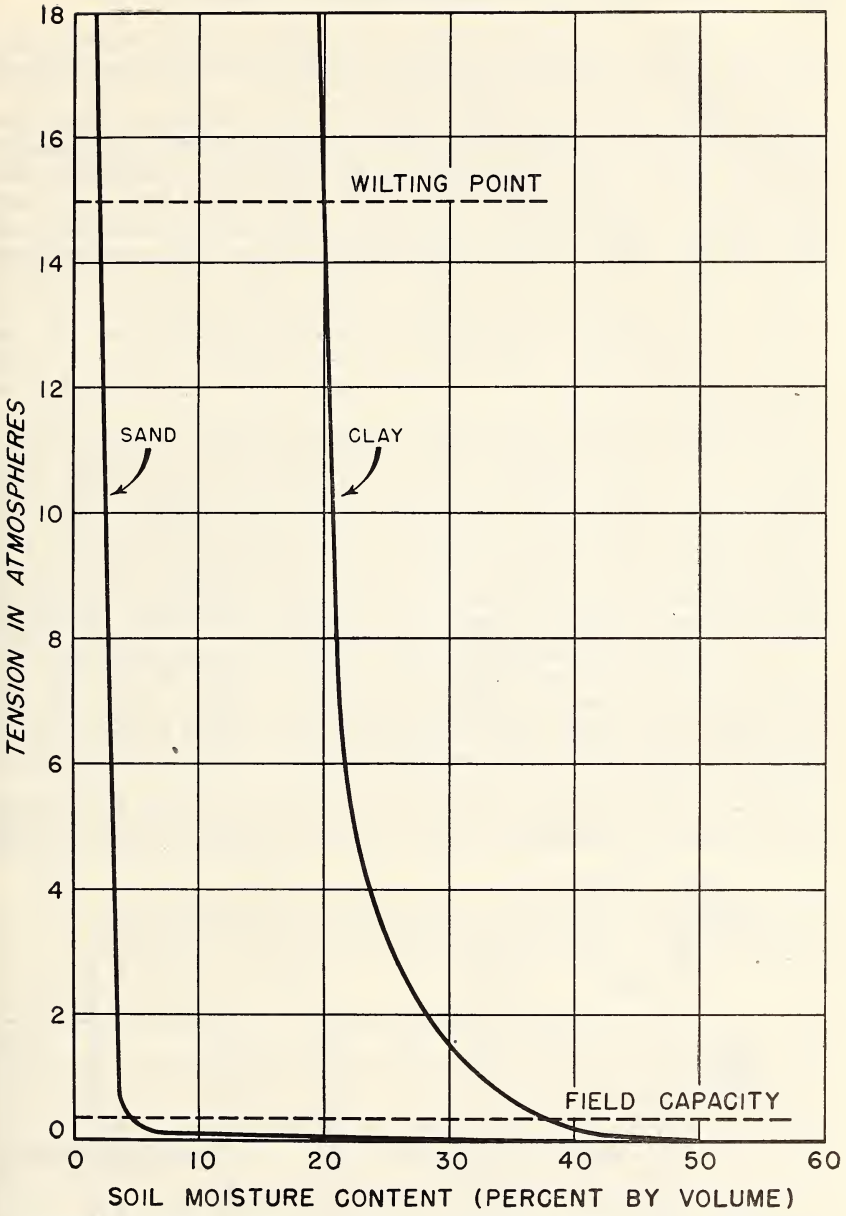

FIGURE 5.—Typical relation between moisture tension and soil-moisture content for sand and clay.

approximately one-third of an atmosphere (about 5 pounds per square inch). At wilting point it is 15 atmospheres (about 225 pounds per square inch). This shows either that vegetation cannot exert a force much greater than this, or that the rate of moisture movement is too slow to satisfy transpiration require-

ments. It must be stressed that the values of field capacity and wilting point, respectively, when expressed in terms of energy, are the same for all soil textures.

However, although energy values are constant at these two points, moisture contents are not. At wilting point, the moisture content of the clay is relatively high because, as explained previously, its small particles give it a large surface area for water adsorption. By contrast, the sand, with larger particles and less surface area, can hold less water at the same energy level. Thus a sandy soil can be dried out more than a clay. Clay has a higher field capacity, however. Because the difference—18 percent—in moisture content of the two soils at wilting point is less than the difference—33 percent—at field capacity (fig. 5), more water is held by the clay in retention storage and more will be available for plants.

To make this point clear, the retention storage of a sand and a clay soil may be compared with the storage of two different sized tanks from which water is removed by two pumps with equal lengths of suction pipe simulating roots of vegetation (fig. 6). The tanks are the same in depth but of different diameter. The similarity between the pumping unit and vegetation lies in the following facts: (1) The equal suction in each tank is comparable to the equal drain on water by vegetation from the two soils; (2) the depths to which the pipes extend are equal and therefore comparable to the wilting point limit in terms of energy, and (3) the upper water levels simulate the energy limit of field capacity in the soil. The energy limits are equal for both tanks both at wilting point and field capacity, as in the case of soils of different texture.

The water removed between the upper level and the bottom of the pipe is comparable to the volume of water removed in the

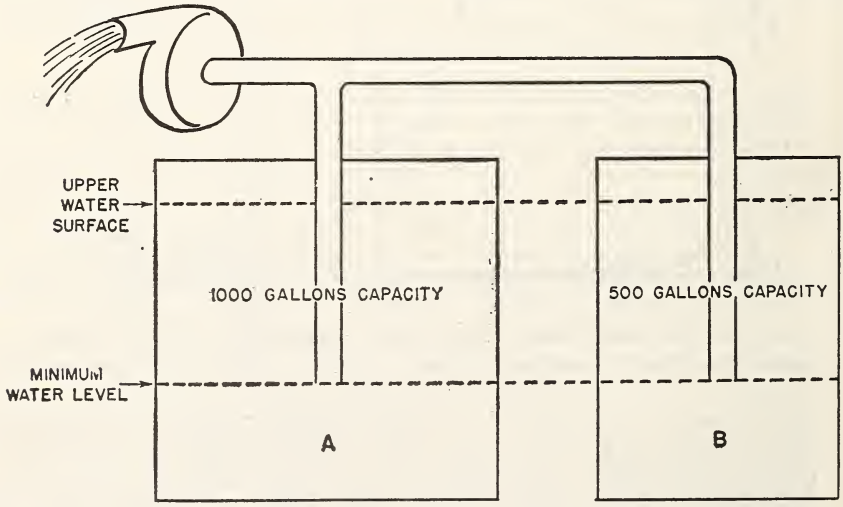

FIGURE 6.—Mechanistic example of the relation between moisture tension and moisture content.

soil between field capacity and wilting point. In each case, the total volume of water removed depends on the volume between the two limits. Thus, while water is lowered in the tanks and in the soil to equal levels of energy, more water is removed from the larger tank for the same reason that more water is removed from the clay. Therefore, in this comparison the limits of energy and moisture content are defined by the dimensions of the tank: the height of the water level shows energy limits, and the tank diameters reflect moisture contents.

Extraction of Moisture by Roots

Roots obtain most of their moisture from soil particles with which they are in direct contact. Although there is some moisture movement from adjacent particles, moisture held by a soil particle several inches from a root does not, in general, become readily available to it. At this moisture level, i. e., where water is being held against the force of gravity, movement from a high to a low zone of moisture is so slow that for all practical purposes it can be ignored. Veihmeyer and Hendrickson (*37*), for instance, found that it took 139 days for water in a soil wetted to field capacity to move 8 inches through a dry soil.

Considering the slow rate of water movement in retention storage, it is apparent that in upland areas where the water table is below the reach of roots the soil moisture available to a plant is confined to that available in the mass of soil encompassed by its root system. Moisture extraction will therefore be limited to the soil volume occupied by the roots. (In areas where roots can draw water from the water table, additional supplies are available.)

When water is available above the wilting point, roots extract moisture simultaneously at all depths in the soil they occupy. The proper understanding of this fundamental fact will greatly clarify the relation of land use to retention storage and related aspects of land hydrology.

Typical moisture extraction curves are shown in figures 7 and 8. The first three examples (figs. 7, *A*, *B* and 8, *A*) show the march of soil-moisture curves for peach trees, desert vegetation, and sugar beets, by successive 1-foot or 6-inch depths. The slopes of the curves in each set appear very nearly parallel, indicating that moisture was being removed at about the same rate throughout the depth of root penetration. Figure 8, *B*, which is derived by plotting together the curves in figure 8, *A*, indicates that for each interval of time there is a decrease in moisture content at each foot of depth. Increases and decreases of moisture content in the upper soil layers during certain months are probably due to small rainstorms, which wetted the upper layers alone, and to the effects of additional moisture losses by evaporation. These charts permit the inference that moisture deficiencies in a given soil mantle at any time will extend to greater depths under vegetation with deep roots than under shallow-rooted vegetation.

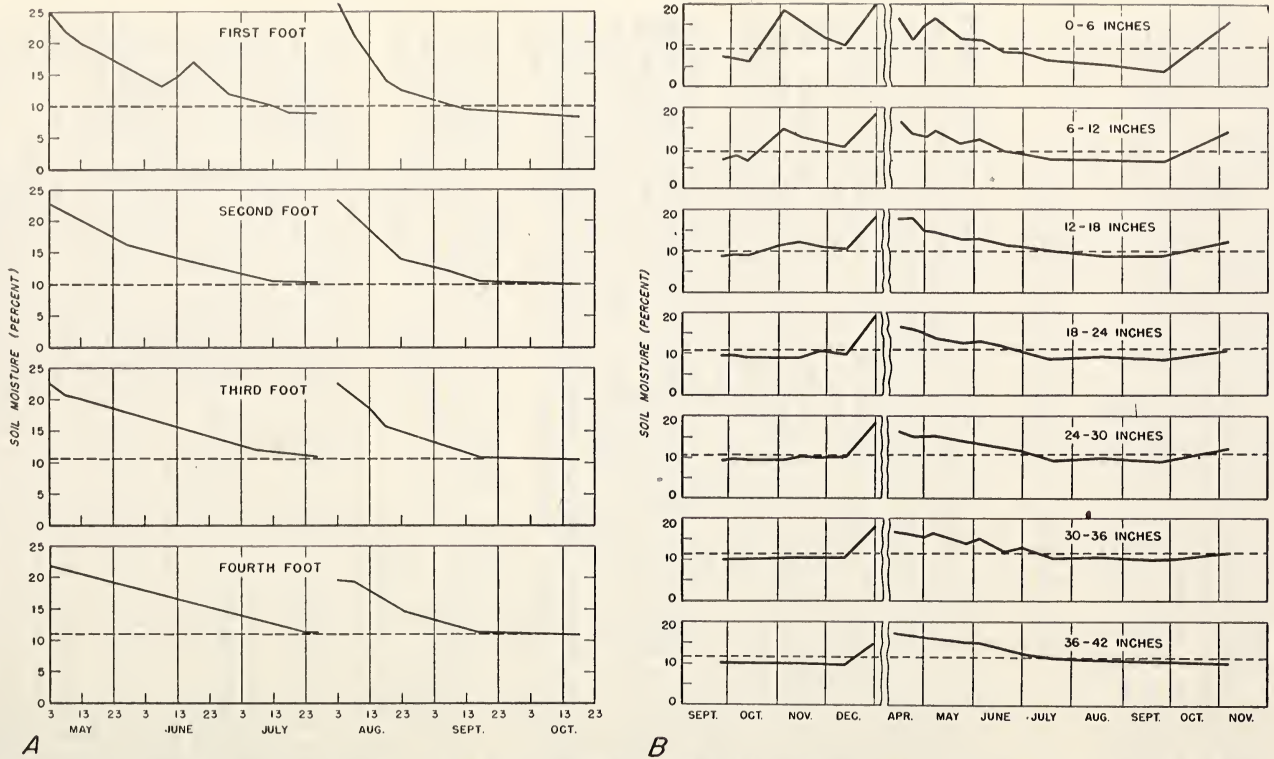

FIGURE 7.—Soil-moisture extraction curves. Permanent wilting percentages are indicated by broken lines (*21*). *A*, Peach trees on Madera clay loam. The increase in soil moisture in June was due to rain, that in early August to irrigation. *B*, Native desert vegetation on Hugo loam. Increases in soil moisture are due to fall and winter rains. Part of the record during the winter months is omitted.

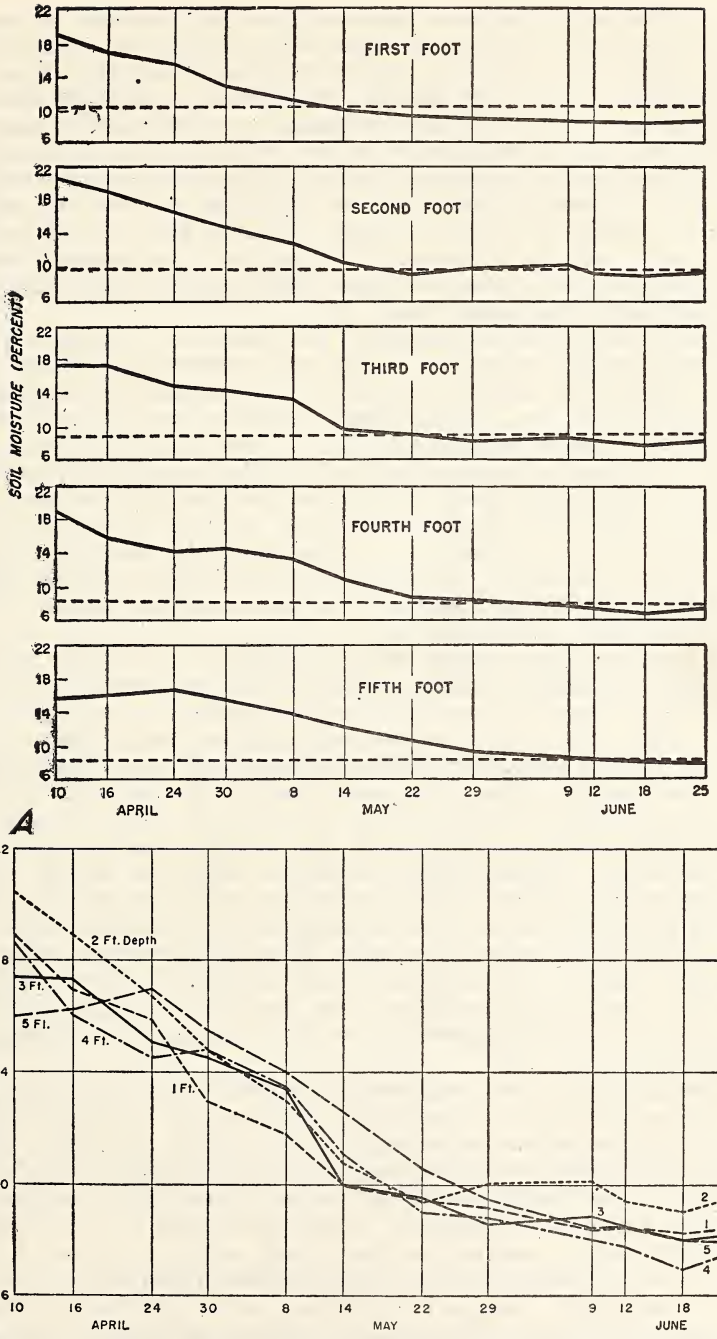

FIGURE 8.—*A*, Soil moisture extraction curves obtained with sugar beets on Yolo loam. Permanent wilting percentages are indicated by the broken lines (*21*). *B*, The curves in *A* superimposed.

The curves also illustrate the principle that the rate of moisture extraction is a function of moisture content; the greater the content, the more rapidly is water removed. From the shape of these curves, as shown in figure 7, it is evident that the rate at which roots extract moisture decreases as soil-moisture content is decreased, or, in relation to energy (fig. 5), when increased force is required to overcome the attraction of moisture to the soil particle. When moisture becomes available at a certain depth, as at the surface following a light rain, the rate of extraction at this depth will be increased. During this time interval moisture extraction will continue at other depths, but at comparatively higher tensions and lower rates.

The foregoing analysis indicates the operation of a governing relationship which tends to reduce the difference between the magnitudes of the forces involved in the extraction of moisture at the various depths. Thus at high moisture contents, soil moisture will be reduced more rapidly than at lower moisture contents, tending to bring the energy levels—and the moisture contents—together. Simply expressed, the higher the moisture content the faster the loss. This is exemplified by the initial recession of the 1-foot curve in figure 7, *A*. The moisture content at this depth at the beginning of the record was greater in the upper foot than at lower depths. But this curve rapidly drops to the relative position of the others.

The relations just stated make it possible to develop a generalization applicable to that portion of the soil mantle which is occupied by roots below the depth affected by evaporation; namely, that this portion of the soil will dry out at rates such that the wilting point—or any other point of equal force—will tend to be reached at about the same time at all depths. If, during this drying period, rainfall wets only an upper layer, the rate of moisture loss will be greater for the wetted portion than for the drier lower depths, thus tending to bring the entire depth to the same energy level. To take an extreme case, moisture extraction from each layer of a soil composed of alternate layers of sand and clay will proceed at rates determined by the shape of applicable tension curves. Then (after starting at field capacity in both layers of soil) whenever the force is at some point between 15 and $\frac{1}{3}$ atmospheres in sand, it will approach the same value in clay.

Hydrologically, these relations are highly significant. Vegetation whose roots are 4 feet deep will provide twice as much opportunity for retention storage in the soil as vegetation on a similar area whose roots are only 2 feet deep. Assuming that 2 inches of water per foot of soil can be stored for retention, a rain falling after wilting point is reached would have to satisfy the 8-inch deficit created by the 4-foot rooting system, but a deficit of only 4 inches for the 2-foot system. Ignoring the briefly effective detention storage for the sake of simplicity, it follows that a rain between 4 and 8 inches would exceed available storage in the shallower-rooted area. No excess would occur in the area occupied by the deeper-rooted plants.

Similar reasoning applies to any intermediate moisture conditions. Rain occurring when moisture content had dropped to 1 inch per foot below field capacity would have to satisfy a deficiency of 2 inches in the shallow-rooted area and 4 inches in the deep-rooted area before it could produce runoff. Following rainfall, additional storage space would be made available twice as rapidly in the deep as in the shallow-rooted area. The rate of drying per unit depth would be the same in both cases, but the soil would be dried to a 4-foot depth in one area and a 2-foot depth in the other.

These considerations point to some salient distinctions between transpiration and evaporation, particularly in connection with the manner in which they create available retention storage. One is that whereas transpiration removes water simultaneously throughout the entire depth occupied by roots, evaporation proceeds downward from the surface. Another is that by acting on a greater volume of soil, transpiration removes a greater amount of water in a unit of time than does evaporation.

Given a soil in which evaporation and transpiration reach to equal depths, a comparison of water loss from a bare area (where water is removed only by evaporation) with that from a well-vegetated area (where transpiration is the principal agent) will show that water is removed more rapidly—that is, retention storage opportunity is created more rapidly—in the vegetated soil. But assuming no addition of moisture to either area, evaporation will eventually remove a greater amount of water because the evaporation process is not governed by the physiological factors which limit transpiration losses.

Evaporation and transpiration do not reach to equal depths in most upland soils. As pointed out previously, unless the soil is loose in structure or cracked, water loss by evaporation is generally limited to the first foot. Transpiration can generally dry out the soil to a greater depth, as determined by the depth and extent of the roots. For example, during a prolonged drought at Rothamsted no water was lost from a bare plot below 18 inches; in a neighboring plot, however, a crop of barley had reduced the water content to a depth of 45 to 54 inches (*35*).

It must be kept in mind, however, that comparisons between evaporation from a bare soil and transpiration serve merely to indicate the manner and magnitudes of water losses. Actually, large expanses of completely bare areas—where only evaporation operates—are uncommon. Where they do occur, as in arid regions, lack of water severely limits evaporation losses. Bare spots within sparsely vegetated areas have a hydrologic importance that is related more directly to the rate at which water can enter the soil surface than to evaporation losses.

Transpiration and Root Characteristics

Since the retention storage opportunity in the soil depends largely upon the extraction of moisture by transpiration, it is apparent that changes in forest and range vegetation by land

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use will affect transpiration and, in turn, the availability of water storage space in the soil. A clue to these relationships lies in the natural behavior of roots. The depth of roots is influenced both by the inherent characteristics of plants and by the environmental conditions under which they exist. That large differences occur among plants with reference to root depth is illustrated by the following data for various tree species in the shelterbelt region.

Root depth of trees in the shelterbelt zone of the Great Plains (22)

Shallow rooted, 1 to 5 ft.	*Intermediate, 5 to 10 ft.*	*Deep rooted, 10 to 20 ft.*
Jack pine	Green ash	Ponderosa pine
Scotch pine	American elm	Hackberry
Norway spruce	Redcedar	Honeylocust
White willow	Russian-olive	Bur oak
Cottonwood	Caragana	Mulberry
Catalpa	Boxelder	Osage-orange
	Black locust	

If storage varied directly with root depth, and the average depth for each group was 2, 8, and 16 feet respectively, the areas with the deepest roots would have 8 times more potential storage capacity than the shallow rooted areas. Figure 9 shows roots of prairie vegetation varying from 1 to nearly 8 feet deep. On this basis the storage space under an acre of mulesears wyethia would be well over six times greater than that under an acre of Sandberg bluegrass (*Poa secunda*).

FIGURE 9.—Variation in depth and distribution of root systems of some important prairie plants (*39*). *s, Sieversia ciliata; w, Wyethia amplexicaulis; ll, Lupinus leucophyllus; lo, Lupinus ornatus; p, Poa secunda; e, Leptotaenia multifida; a, Agropyron spicatum.*

The effects on soil storage of reducing transpiration by reducing root depth have been demonstrated. In one experiment in Utah (*9*), deep-rooted aspen trees were felled, leaving the shallower-rooted herbaceous plants. As less water was used by the latter, a larger proportion of the precipitation became available for stream flow.

Another illustration of the relation of root depth to storage is provided by a study in southeastern Washington (*24*). The soil was a silty clay loam at least 17 feet deep. Soil moisture changes were measured respectively on land in summer fallow (bare), bunchgrass, alfalfa, and a plantation of mixed conifer and hardwood species. If it is considered that the depth to which moisture is extracted closely approximates the depth of root penetration, then the differences in rooting depth and in soil moisture storage by the end of the growing season may be evaluated. On this basis, the greatest storage, 48 inches, occurred under the alfalfa cover, whose roots extended apparently to a depth of at least 14 feet. The next greatest, 41 inches, was under the tree cover, whose roots were apparently 12 feet deep. The storage under bunchgrass, whose roots apparently went only 6 feet down, was 20 inches. The least amount of storage was found under the bare area. (The moisture content in inches was derived on the basis of an estimated storage capacity of 30.9 inches per 9 feet of soil as given in the publication cited.)

The inherent root characteristics of plants do not appear to limit plant occupancy only to those areas where maximum development can occur. Most plants will live even if their roots are prevented from reaching their maximum depth by such obstacles as poor aeration or an impermeable layer. Within certain limitations they have the ability to adjust themselves to soils shallower than the optimum for best root development. For example, the shallow Volusia soils of New York support aspen, beech, birch, and maple whose roots fully occupy the 12 to 18 inches above the impermeable layer. Yet, in the neighboring Lordstown soil, roots of the same species penetrate to within an inch or two of bedrock, which here lies 30 to 36 inches beneath the surface. Aspen growing on deep soils on the Wasatch Front in Utah sends its roots down at least 6 feet.

This ability of plants to occupy soils shallower than their inherent root depth capacity is highly important in watershed management. Inherently deep-rooted plants growing on shallow soils will take advantage of any increases in soil depth resulting from the incorporation of organic matter. In this way they will be instrumental in increasing the total amount of moisture available for transpiration.

In addition to differences in transpiration associated with varying root characteristics, there also appear to be marked differences in transpiration rates among different species during the growing season. These differences, however, are believed to have little significance for watershed management. Based usually on experiments during which the species studied were provided with as much water as they could use, they have little application

to actual field conditions where the amount of water available is the over-all limiting factor to transpiration.

Thornthwaite's curves of potential evapo-transpiration and water supply indicate that for most sections of the country there is not sufficient rainfall during the growing season to meet potential evapo-transpiration rates (*36*). Therefore, disregarding differences in root depth, it is extremely doubtful whether in any one locality one type of forest vegetation will use greater amounts of water than another type in equal periods of time during the growing season. Given all the water needed, differences would probably exist, but in upland areas this condition does not apply, so that differences in transpiration can for the most part be ascribed to differences in soil depth occupied by plant roots.

It is possible that different species or types of upland vegetation may transpire at different rates during winter periods of slow growth and relatively great moisture availability. Differences may also be related to the varying times when growth begins and transpiration rates increase in the spring. Whether winter transpiration is greater for broadleaf trees or conifers has not yet been satisfactorily determined. Some studies show little difference (*34*). The time that spring growth begins may perhaps be more significant. Thus, Hoover [5] has found that in the spring, loblolly pine begins to draw on water in retention storage about 2 weeks before broomsedge. Differences among deciduous species in the time required to develop to full leafing in the spring must also affect transpiration rates.

Transpiration is also affected by the rate of development of the root system during the growing season, particularly in the case of annual plants. Here the growth of roots provides increasing opportunity for water removal as the season progresses. The rates of water extraction for various depths (*11*) are illustrated, as shown by Kramer (*29*), in figure 10. Roots of the tomato plant reached the 4-foot level on June 15, 6 feet on July 20, 8 feet by about August 15, and 12 feet on September 3. Very nearly all the available water was used throughout the depths measured. The figure also indicates that 1.7 inches of storage capacity was available on June 15 in the upper 2 feet of soil. By October 23, approximately 21.0 inches of storage space was available throughout the 12-foot depth, or about 12 times as much. This same relationship between water loss and root development would also hold for the growth of perennial vegetation.

Effects of Vegetative Treatment on Transpiration

Since available storage space depends largely on transpiration losses, any change in the condition of vegetation that alters transpiration rates will also affect storage opportunity. Thus, the removal or killing of vegetation affects transpiration. Leaf area and hence the transpiring surface is reduced. Where only part

[5] Unpublished data, Calhoun Experimental Forest. 1950. U. S. Forest Serv., Southeast. Forest Expt. Sta.

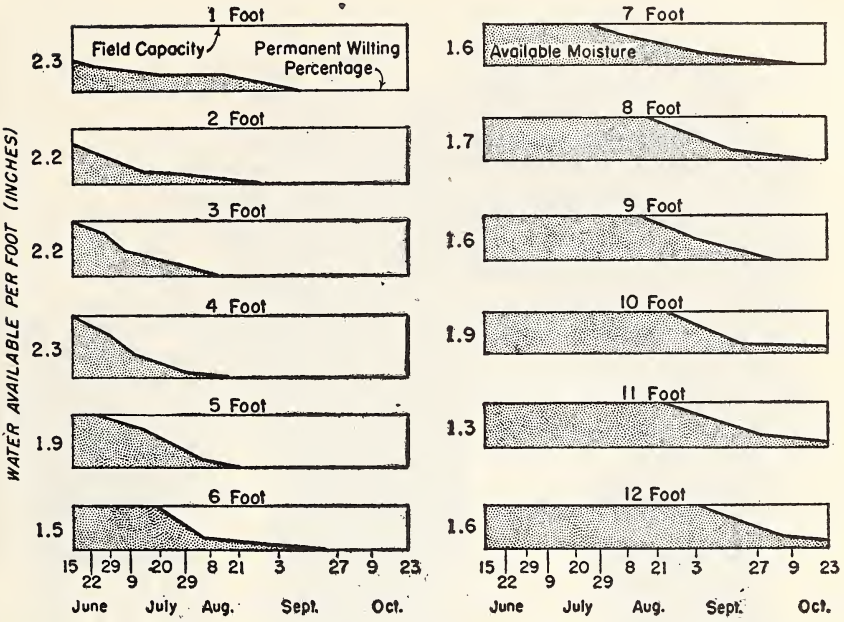

FIGURE 10.—Rate of water extraction by tomato plants from various depths in an unirrigated soil. The clear areas in each rectangle indicate the proportion of available water removed; the shaded areas signify water still available for plant use (*11*).

of a plant is destroyed, transpiration may or may not be reduced, depending on whether the rate can be maintained with the remaining leaf surface. However, since root growth depends on the plant food produced by photosynthetic activity in the crown, any reduction in leaf area can be expected to reduce root growth. Thus there tends to be a balance between crown and root activity. Killing of vegetation reduces root depth and in consequence, the depth of soil from which roots extract moisture.

This puts an entirely new aspect on the effects of forest fire on the functioning of a watershed. An intense fire may decidedly change the rate at which moisture is extracted from the soil. Where the destruction of the existing stand is followed by an invasion of different plant species, the rooting systems of the invader may not extend as deeply as those of the original vegetation. Then, the opportunity to extract water from the soil will be lessened, more moisture will remain in the soil, and therefore retention storage opportunity will be decreased. Similarly, clear-cutting a forest may have a marked hydrologic effect where the shrubs, and especially the herbaceous growth which comprise the understory, have shallower roots than the trees. Should the former tree species become reestablished, the former hydrologic conditions would be expected to return.

When an area is clear-cut and immediately replanted, much time may elapse before the growing roots fully occupy the soil and thereby provide the maximum space for water storage.

FIGURE 11.—Plants of little bluestem grown from sods of similar size taken from an excellent, a good, and a poor pasture. This is the new growth after 6 weeks. Weight of tops was 30.22, 12.89, and 1.87 gm.; weight of roots was 4.70, 1.72, and 0.33 gm. respectively. Roots were progressively more slender with decrease in vigor (*40*).

Coile (*8*) found that root development of loblolly pine in the surface horizon (1 foot deep) increased rapidly until the stand was 20 to 30 years old. Afterwards the rate was slower but still apparent during the period when climax species (hardwoods) were replacing the pine.

Changes in storage resulting from the forest treatment are indicated by data from the Coweeta Hydrologic Laboratory of the Southeastern Forest and Range Experiment Station. By felling all the vegetation on a forested watershed, total water yield the first year was increased by 65 percent. After 10 years of new growth on the watershed, the yield has dropped from 65 percent above normal to about 20 to 25 percent. This experiment indicates that as more water is transpired by the expanding vegetation, more retention storage space is made available and less of the precipitation goes to stream flow.

Grazing also affects root development and thus soil storage capacity. Kramer (*29*) has reviewed the literature on the effect of grass clipping and grazing on root development. He cites among other sources the findings of Biswell and Weaver (*7*) that 4 to 7 clippings of several species of grasses reduce the dry weight of the roots to an average of 10 percent of that of the unclipped controls; the findings of Weaver and Darland (*40*) that overgrazed plants develop very few new roots as compared with ungrazed plants, and further, that many of the old roots die (fig. 11). Thus retention storage opportunity in a grazed area—especially one subject to overgrazing—becomes less than that of an ungrazed area.

Retention Storage and Stream Flow

The previous discussion has been limited entirely to water in retention storage. Three facts have been stressed: First, the movement of this water is so slow as to be practically nonexistent, so that water available to plants must come from within the mass of soil occupied by the root system; second, the volume of retention storage is affected by land-use practices which affect the soil depth and its water-holding capacity through the production and decomposition of organic matter; and third, the vegetal conditions resulting from land-use practices determine retention storage opportunity in accordance with root characteristics.

The effects of treatment of vegetation on retention storage are reflected in the amount and rate of stream flow. By managing forest and range vegetation so that both the depth to which water is stored and the water-holding capacity of the soil particles are increased (thus increasing total storage capacity), more rainfall can be stored. Consequently, less will drain off to streams and water yield will be reduced. In addition, it may be possible to increase the total retention storage capacity of deep soils by encouraging or introducing deeper rooted vegetation. By drawing water from greater depths, such vegetation would provide greater space for storage of precipitation so that less would go to stream flow.

Watershed treatment not only affects total retention storage, but also the opportunity for retention storage at any one time. In the absence of vegetation, available storage space is created solely by evaporation losses. Under such conditions, however, the

available storage space often cannot be utilized because the rate of entry of water into the soil may be greatly restricted by a surface only slowly permeable to water. On the other hand, where vegetation is present, evaporation is reduced, but the total loss by transpiration plus evaporation is commonly greater than the loss by evaporation alone from a bare area. Thus retention storage opportunity is not only greater under vegetation, but is also created more rapidly than in bare areas.

In summary, retention storage opportunity in a given area may be increased in the following ways: (1) Maintaining a stand of vegetation whose roots fully occupy the soil so that maximum storage space is provided; (2) encouraging the growth of deep-rooted vegetation; (3) favoring those forms of plant growth having the longest period of water use.

The value of having an ample volume of retention storage opportunity as a means of reducing the possibilities of floods becomes especially significant in periods of heavy rainfall during the growing season. The amount of space at such times will depend to a large extent on how much water has been removed by vegetation since the previous rain. This, in turn, depends on its water-using characteristics. Deep-rooted growth will have removed the most water; hence, flood flows will be reduced more by such growth.

The opportunity for retention storage may also be reduced by watershed treatment. Here the objective would be to reduce transpiration loss, thus making less storage available for precipitation, and more precipitation available for stream flow. A prime requirement of such an operation would be the maintenance of soil stability.

The effects of vegetation and its treatment on stream flow have thus far been described mainly in volumetric terms. Over and beyond this aspect, however, is the effect on water movement in the soil and the temporary (detention) storage that occurs during this movement. By influencing water movement and detention storage, treatment of vegetation plays an important part in determining the apportionment of precipitation to surface flow, subsurface flow, and base flow, respectively. Because each type of flow has a characteristic rate, plant-cover treatment alters the length of time required for precipitation (other than that which falls directly on and along the streams) to reach channels. In this manner it directly influences stream flow rates as reflected in the shape of the hydrograph.

SOIL-WATER MOVEMENT AND TEMPORARY STORAGE

The rates at which water enters, moves through, and is temporarily held back in the soil are determined by the size, number, and arrangement of the large (noncapillary) pores. The temporary delay which occurs during water movement is termed "detention" storage, as distinguished from the "retention" storage in the small (capillary) pores.

Factors Affecting Soil Porosity

Noncapillary porosity and water movement and detention storage are affected by a number of factors, of which soil texture will be considered first. Assuming for the time being that soils are composed of single-grain particles only, the rate at which water moves between them will depend on their size. As previously noted in the discussion of retention storage, size of particles (soil texture) is important because of its effect on surface area. Small particles provide greater resistance to water flow than large ones because the water comes in contact with more surface, and greater friction is developed.

This is similar to a phenomenon that occurs in pipe flow. Under the same pressure more water will flow through a large pipe of a given cross-section area than through a number of small pipes with the same total area. The reason is that more surface is in contact with the water in the smaller pipes than in the single large pipe, hence total friction and resistance to flow is greater in the small pipes. For instance, according to the equation for discharge of water through a pipe, discharge is directly proportional to the fourth power of its internal radius. On this basis, all other things being equal, it would require 16 pipes, 2 inches in diameter to furnish as much water as one pipe 4 inches in diameter.

The effect of texture on rate of water movement (permeability) is very evident in the range of soil textures from coarse sand to very fine sand. After the introduction of clay particles in sufficient quantities to change the texture to a clayey soil, permeability becomes negligible and the effect of further changes in texture is not so apparent. Nevertheless, rates of several inches per hour have been reported for clays. This marked difference is due to the effect of the admixture of organic matter and clay on the size and distribution of the noncapillary pores in the more permeable soil.

Adding organic matter to the soil increases its noncapillary pore space by separating the soil particles and creating additional large pores. The organic matter itself has a light volume weight; peat, for instance, has a weight of 0.4 per unit volume. Associated with its light weight is high porosity, so that water moves rapidly through it. Thus the incorporation of organic matter into the soil reduces its weight. Less weight per volume in turn indicates greater pore space, largely of a noncapillary nature.

Besides producing a physical separation of the particles and a characteristically high noncapillary porosity, organic matter also increases noncapillary porosity by the formation of aggregates. In this process organic matter functions as a binding agent, the soluble organic compounds forming a coating over the aggregation of particles. An additional requisite of aggregate formation is the presence of some clay particles, a clay component of at least 8 to 10 percent being required (*26*).

The binding together of the particles tends to concentrate the pore space into large openings, thus increasing noncapillary

porosity. Accompanying this process is an increase in total porosity, and, under some circumstances, as previously noted in the discussion of retention storage, a decrease in capillary porosity.

Aggregation results in a reduction in the number of separates and hence in their specific surface in the original soil volume, and this in turn reduces retention storage capacity. As further organic matter is incorporated and as aggregation proceeds accordingly, the mass of soil expands upward (the line of least resistance to pressures created by aggregation) producing in effect a reduction in its volume weight. The over-all result is that the former retention capacity for the entire mass involved may be restored or even increased.

The larger the aggregates formed, the greater the amount of noncapillary pore space. For example, a soil with particles less than 0.5 mm. in diameter had a total porosity of 47.5 percent, of which 2.7 was noncapillary porosity and 44.8 capillary porosity. Aggregation of particles into 1 to 2 mm. aggregates increased total porosity to 54.7 percent. Noncapillary porosity increased to 29.6 percent or about 11 times greater than the original 2.7 percent. Capillary porosity decreased to 25.1 percent. When aggregates increased to sizes ranging from 3 to 5 mm., total porosity increased to 62.6 percent, and the noncapillary porosity to 38.7 percent, while the capillary porosity decreased to 23.9 percent. Increasing the noncapillary porosity from 2.7 to 38.7 percent made detention storage more than 14 times greater. There also was a 32-percent reduction in volume weight, equivalent to a 32-percent increase in soil depth (*5*).

It becomes obvious therefore that aggregation can so change the porosity of a soil as to mask completely the effect of texture on porosity. In the foregoing example the original soil was a clay. After aggregation, a textural analysis would show that it was still a clay, but so far as noncapillary storage space is concerned, and water movement as well, this soil more nearly resembles a sand. Texture thus becomes a limiting factor to water movement only in soils where relatively little organic matter is present, as in agricultural soils, or in subsoils below the depth of accumulated organic matter.

Since organic matter is the principal determinant of noncapillary porosity, noncapillary porosity will decrease with soil depth because the organic content also decreases with depth. Figure 12 illustrates typical differences between a forest soil and that of an abandoned field (*27*). Noncapillary pore space decreased with depth in both instances. In the forest soil, however, it remained considerably higher than in the agricultural soil to the 18-inch depth. Below that depth pore space was the same for both soils. Thus in the upper 6 inches of the forest soil noncapillary pore space was 30 percent of the total volume, in the next 12 inches, 12 percent, and the next 18 inches, 3 percent. In the agricultural soil, however, noncapillary pore space was only 10 percent in the upper 2 inches of soil depth and 7 percent in the next 4 inches. Below 6 inches and to a total depth of 36 inches, non-

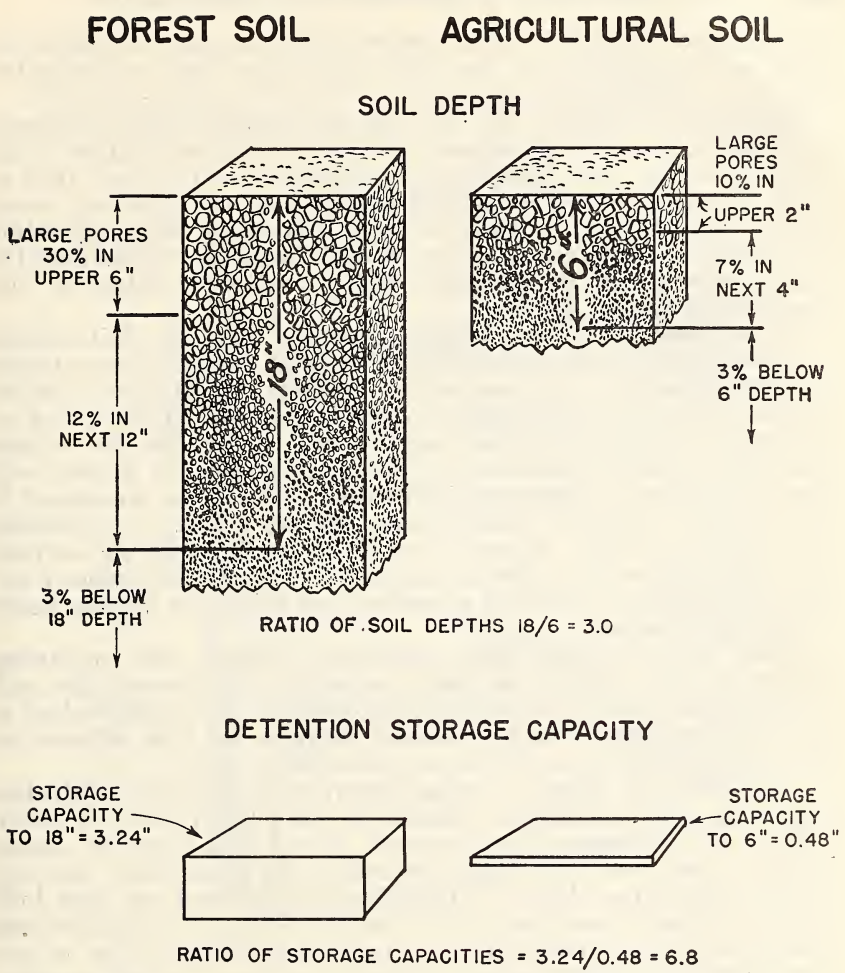
FOREST SOIL
AGRICULTURAL SOIL
SOIL DEPTH
LARGE PORES 30% IN UPPER 6"
12% IN NEXT 12"
3% BELOW 18" DEPTH
18"
LARGE PORES 10% IN UPPER 2"
7% IN NEXT 4"
3% BELOW 6" DEPTH
6"
RATIO OF SOIL DEPTHS 18/6 = 3.0
DETENTION STORAGE CAPACITY
STORAGE CAPACITY TO 18" = 3.24"
STORAGE CAPACITY TO 6" = 0.48"
RATIO OF STORAGE CAPACITIES = 3.24/0.48 = 6.8

FIGURE 12.—Comparative porosities, depths, and water-holding (detention) capacities of a forest and an agricultural soil.

capillary porosity was 3 percent. The effect of organic matter on noncapillary porosity extended in the forest soil to a depth 3 times as great as that in the agricultural soil, but as shown in the lower part of figure 12, its influence on detention storage capacity throughout these depths was even greater.

INFILTRATION AND PERCOLATION

Noncapillary pores represent the avenues of free water movement in the soil. Subsurface flow moves laterally through them and water which becomes base flow moves downward through them on its way to underground reservoirs. In short, the noncapillary pores conduct the water to the stream. The effects of plant-cover conditions and treatment on these pores determine

to a large extent how water moves to the stream—whether by surface, subsurface, or base flow—and how fast, as indicated by the stream-flow hydrograph.

Water movement in the soil can be separated into two more or less independent processes, infiltration and percolation. Infiltration refers to the entry of water into the surface half inch of soil. Percolation refers to the rate of flow through any lower level. A distinction is made between infiltration and percolation because certain forces—such as rainfall impact—which affect the soil surface, and consequently water movement rates, do not operate at deeper levels.

The maximum rate at which water can enter the soil surface is here defined as the *infiltration capacity* of the soil. This capacity is determined by the noncapillary porosity of the soil surface after it has been thoroughly wetted, but not to the point of saturation. Other conditions being equal, the greater the volume and the larger the diameter of the large pores, the greater will be the infiltration capacity. Infiltration capacity is considered to be strictly a surface phenomenon and as such is independent of other hydrologic processes which occur beneath the surface. As defined here it is affected neither by percolation capacity nor by the amount of water in retention and detention storage below the surface layer.

The rate at which water can move through the subsurface layer is here defined as the *percolation capacity* of the soil. Percolation capacity is therefore considered to be independent of infiltration capacity and soil-water storage. Its rate depends on the noncapillary porosity of the subsurface layer.

As indicated above, a major distinction between infiltration capacity and percolation capacity is that the former is determined by the noncapillary porosity of one thin layer of surface soil, whereas the latter is determined by the noncapillary porosity of the subsurface layer. It has also been pointed out that infiltration capacity operates only to the point where the soil becomes saturated. Once this point has been reached—as after a prolonged rain has filled the large pores throughout the soil mantle—the rate at which additional rain can enter the soil is determined by the rate at which gravitational water in the soil can move out as subsurface flow, and downward as deep seepage. Thus, no matter how high the infiltration capacity may be, once a soil has become saturated, the rate at which further infiltration can occur will correspond with the rate or rates at which water flows downward and outward. The purpose of distinguishing between infiltration *capacity* and infiltration *rate* is to isolate the conditions, described later, which determine the rate of water entry into the soil.

These processes are illustrated in figure 13 which represents a mechanistic interpretation of the processes of infiltration, percolation, and storage.

The tank has a total storage capacity of 1,500 gallons. An outlet (*P*) is provided one-third of the distance from the top, with a discharge capacity of 2 gallons per minute when the tank is full.

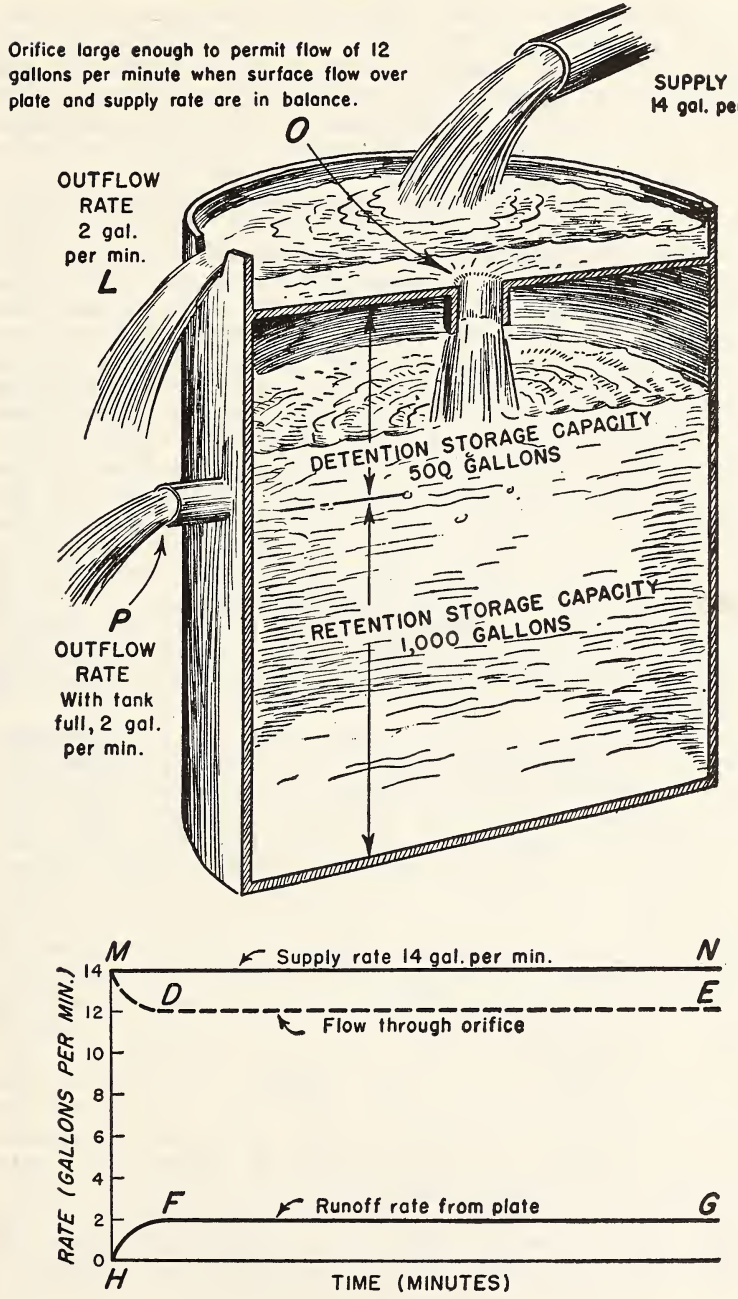

FIGURE 13.—Mechanistic example of infiltration, percolation, storage, and surface runoff.

A rimmed steel plate with an orifice (*O*) closes the top of the tank. A lip (*L*) is cut in the rim so that 1 inch of water can accumulate on the top. Water is supplied to the surface of the plate at the rate of 14 gallons per minute. Before flow can occur over the lip (*L*) a depth of 1 inch must be built up on the plate. When this occurs, a discharge of 12 g.p.m. through the orifice is assumed, leaving 2 g.p.m. to flow through the lip.

The analogy is as follows. The supply rate simulates rainfall, the water passing the lip (*L*) surface runoff, the water going into the orifice (*O*) infiltration, and that discharged through (*P*), percolation. Rates of flow are shown in the lower part of the figure. Line M–N represents the supply rate. Rate of flow over the lip (*L*) is shown by line H–F–G, H–F being the time required for storage on the plate to build up to produce a flow of 2 g. p. m. through the lip (*L*) and 12 g. p. m. (infiltration capacity) through the orifice (*O*). If there were no way to measure the discharge through the orifice, it could be calculated by subtracting plate runoff from the supply rate, resulting in line M–D–E, which shows the rate of discharge through (*O*). At the beginning, however, the portion M–D of the curve M–D–E represents both the flow rate through the orifice and the rate of storage on the plate surface. The water stored on top of the tank during time M–D simulates depression storage on the soil surface. (Curve M–D–E represents the flow through (*O*) only up to the time when the tank is filled. It does not show the drop in the flow through (*O*)—from 12 g. p. m. to 2 g. p. m.—after the tank is filled.)

The water flowing across the plate through (*L*) represents surface runoff, that going through (*O*) represents infiltration, and the water first entering the tank for storage below outlet (*P*) represents capillary water. The 1,000-gallon storage space represents retention storage space, and, as in the soil, no flow can occur through the outlet (*P*) until this space is filled. Thereafter, subsurface discharge by percolation will occur through outlet (*P*) but, since percolation capacity (2 g. p. m.) is less than the supply rate (12 g. p. m.) through the orifice, water will continue to be stored in the tank. This storage represents noncapillary or detention storage. When detention storage is satisfied the tank is full, and flow through the orifice (*O*) will be reduced to the 2-g. p. m. capacity of the subsurface discharge at (*P*). Surface flow through (*L*) will then increase to 12 g. p. m. The difference between the surface flow rate (12 g. p. m.) and the supply rate (14 g. p. m.) will no longer reflect the infiltration capacity of the orifice, but rather the rate of outflow at the subsurface outlet (*P*). In other words, the final or sustained infiltration rate at which the water enters the tank is no longer its infiltration capacity because this rate is determined by forces operating below the surface, that is, the percolation capacity.

Factors Affecting Infiltration

Since infiltration capacity is determined by the number and size of noncapillary pores at the soil surface, any activity which

affects these characteristics will influence infiltration capacity. The size of the large pores may be reduced by blocking them with smaller particles or by compacting the soil.

Numerous tests have shown that raindrops falling on a bare soil break down the soil aggregates. As they break they separate into smaller particles, which are washed into the pores of the soil surface by infiltrating rainwater. These smaller particles seal the pores, thereby reducing their size and total volume.

How vegetation reduces the effects of rainfall impact and thus helps maintain soil porosity is indicated by table 1. Differences in total and noncapillary porosity between covered and bare soils show the effects of rainfall impact. For instance, total porosity of the humus calcareous sand with rye cover was 62.2 percent and 58.3 percent without cover. The difference in noncapillary porosity was much greater, 33.5 percent under rye in comparison with 22.1 percent bare, a reduction of approximately one-third. On the other hand, the washing of small particles into the large pores increased the capillary porosity of the bare soil. In this instance it increased from 28.7 to 36.2 percent.

Since vegetation protects the soil against rain impact, thus maintaining the noncapillary porosity of the surface soil and the readiness with which water enters the soil surface, there should be substantial differences between bare and vegetated areas in the depths to which water penetrates. Further, the infiltration capacity of a soil with sparse cover should be expected to increase with an increase in plant density. These differences have been demonstrated. Artificial rainfall applied with an infiltrometer to a limestone clay soil with a 6-percent browse cover penetrated to a depth of 18 inches directly beneath the plants but to a depth of only 4 inches under bare soil.[6] Similarly, in a sandstone loam having a sagebrush cover occupying 37 percent of the surface, moisture penetration under the canopy was 6 inches as against only 1 inch in the bare areas.

TABLE 1.—*Effect of plant cover upon soil porosity* (*5*)

Soil type	Type of cover	Total porosity		Noncapillary porosity		Capillary porosity	
		No cover	Cover	No cover	Cover	No cover	Cover
		Percent	*Percent*	*Percent*	*Percent*	*Percent*	*Percent*
Humus calcareous sand	Rye	58.3	62.2	22.1	33.5	36.2	28.7
	Peas	59.2	63.6	21.0	32.3	38.2	31.3
	Vetch	65.5	69.2	29.0	39.0	36.5	30.2
Quartz sand	Sweet clover	45.4	46.0	34.9	43.9	10.5	2.1
Loam	do	50.9	58.2	31.9	45.3	19.0	12.9

[6] U. S. FOREST SERVICE INTERMOUNTAIN FOREST EXPERIMENT STATION. ORIGINAL RECORDS OF INFILTROMETER RUNS, SEVIER LAKE WATERSHED. 1940.

Infiltration capacity increased on both soils with an increase in cover density. On the limestone clay, infiltration capacity was 1.3 inches per hour when the cover density was 6 percent; with a density of 35 percent, infiltration capacity increased to 2.5 inches per hour. On the loam soil infiltration capacity was only 0.09 inch per hour with a cover density of 3 percent, but with a density of 30 percent, it was 0.16 inch per hour.

Soil compaction is a widespread cause of reduced infiltration capacities and therefore of high surface runoff rates. When a soil is compacted its total porosity is decreased; more important, the reduction affects the noncapillary pore space more than the capillary. Tests showed that the total porosity of the upper inch of an ungrazed clay loam bluegrass pasture was 58.9 percent in contrast to a total porosity of 32.1 percent found in a heavily grazed pasture on the same soil (*1*). Almost all this difference in total porosity was due to the difference in noncapillary porosity values, namely, 33.1 percent for the ungrazed areas, 6.1 percent for the heavily grazed.

Although grazing, walking, or passage of a tractor or truck will usually compact only the upper few inches, in most soils this is sufficient to determine how much rainfall enters and how much runs off. In the above-mentioned pasture tests, a simulated rainfall of 1.4 inches per hour infiltrated into the uncompacted area with no surface runoff. On the compacted area replicated infiltrometer tests with the same rainfall intensity gave infiltration rates of 0.41 and 0.28 inch per hour, with 71 and 80 percent, respectively, of the rainfall running off the surface.

Compaction effectively decreases infiltration rates on both bare and vegetated soils, whereas rainfall impact reduces infiltration only on bare areas. Damaging and lasting compaction of given areas is associated with frequent passage of farm machinery or logging equipment, or by trampling under concentrations of livestock, big game, or persons. This damage is so common that its effect on surface runoff is often overlooked.

In regions subject to freezing and thawing, infiltration and soil permeability may be affected as a result of the formation of certain readily distinguishable types of frost. Such conditions are associated with the type, condition, or treatment of the vegetal cover, as reflected in soil compaction, depletion of surface organic material, or both.

Percolation and Storage

After water has infiltrated into the soil, the rate of its continued downward movement (percolation) depends on the noncapillary porosity of the various soil layers through which it passes. The permeability of a soil (or relative ease with which water passes through) may vary considerably from layer to layer, or remain relatively constant for considerable depths. However it may vary, permeability affects the disposal of precipitation into surface, subsurface, and base flow, and the utilization of storage space.

Percolation capacity of a soil is limited by the least permeable layer with which downward moving water comes in contact. If this layer allows water to pass through at a rate faster than it enters the soil, movement of water through the soil and utilization of storage space will be governed by the surface conditions that determine infiltration capacity. If the percolation capacity of the least permeable layer is less than the infiltration rate, this layer can govern the rate at which water enters the soil, as well as the use of storage space, and therefore the disposal of precipitation.

Figure 14 shows the types of relations between percolation, infiltration, and detention storage that may occur under given conditions. The curves represent the respective cumulative de-

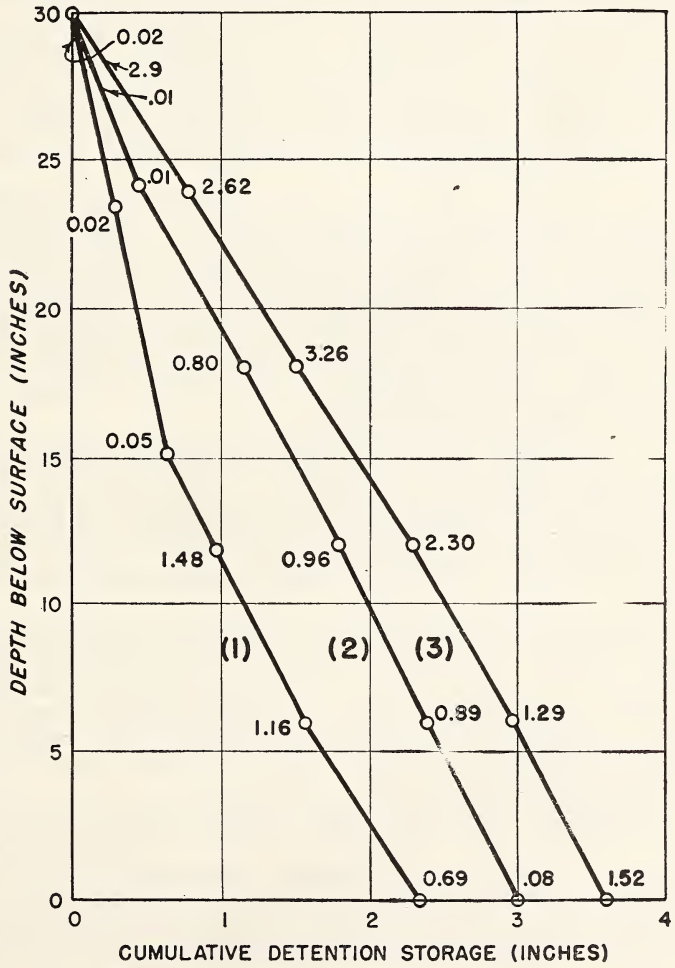

FIGURE 14.—Detention storage of three soils in relation to depth. Infiltration capacities at zero depth, and percolation capacities at other depths, are indicated at plotted points.

tention storage capacities of three soils to a depth of 30 inches. By using such a curve to trace the disposition of a 3-inch rain falling at the rate of 1 inch per hour after retention storage has been satisfied, the processes involved will be made clearer.

Curve *1* represents a soil whose detention storage capacity is 2.35 inches, of which 0.65 inch is below the 15-inch depth. The percolation capacity at the 15-inch depth and below is less than the infiltration capacity at the surface. Based on infiltration capacity, the amount of water that would enter the soil is 3×0.69 inch or 2.07 inches; the rest of the rainfall, or 0.93 inch, would immediately run off the surface. However, the percolation capacity of 0.05 inch per hour at the 15-inch level will permit passage of only 0.15 inch of water in the 3-hour period. This amount, plus the 1.70 inches above the 15-inch level, or a total of 1.85 inches, represents the total storage available for the rainwater. Thus, 1.15 inches must be disposed as surface runoff instead of the 0.93 inch expected when calculated on the basis of infiltration capacity alone. This case illustrates how percolation capacity and detention storage can affect the amount of water that enters the soil.

Curve *2* represents a soil which has greater storage capacity (3.00 inches) than that represented by curve *1*. Yet, because its infiltration capacity (0.08 inch per hour) is less than the percolation capacity at the 18-inch depth and above, even less of its storage capacity is used than in the previous example. At a rate of 0.08 inch per hour only 0.24 inch of rainfall would enter the soil during the 3-hour rain. The rest, 2.76 inches, would run off. As the storage capacity for this soil is 3 inches, 2.76 inches of this space would go unused.

Curve *3* represents a soil with high infiltration capacity and high percolation capacities throughout the 30-inch depth. Since these capacities exceed the rainfall rate, they impose no limitations upon the rates of water entry and movement. Furthermore, as the storage capacity of the soil mantle is 3.6 inches, the entire 3 inchs of rain will be detained. Consequently, no surface runoff will occur.

The foregoing comparison of the differing hydrologic characteristics of three soils indicates the nature and importance of the processes which determine the volume of surface runoff. In the first case, percolation capacity was the limiting factor; in the second, infiltration capacity; and in the third case, each capacity was sufficient to permit the ready entrance and movement of water through the soil.

These examples also show that usability of storage space depends on infiltration and percolation values in relation to rainfall intensities. Although much storage space may be present in a soil, it may remain unused because water cannot move into it rapidly enough. Rainfall intensity is therefore an important factor in determining the usability of storage and also the volume of surface runoff. Had the 3 inches of rain been spread over a period of 6 hours, runoff and use of storage would have been altered in the case of the first two soils. In case 1, use of storage

below the 15-inch depth would have increased from 0.15 inch to 0.30 inch, leaving 0.15 inch less for surface runoff, or reducing its total from 1.15 to 1.00 inch. In case 2, the amount of water entering the soil would have doubled (from 0.24 inch to 0.48 inch) reducing surface runoff from 2.76 inches to 2.52 inches.

If the initial intensity of a storm is 0.69 inch per hour, no surface runoff will occur from the soil represented by curve *1* until usable noncapillary storage is satisfied. Thereafter, even a lesser intensity will produce runoff.

Percolation and Root Development

The effect on percolation of organic matter and aggregate formation is indicated by their influence on noncapillary porosity, as previously described. Root development is another factor which influences percolation. Roots may be instrumental in providing avenues for percolation and storage of water in the subsoil by penetrating below the zone of high organic content. What is known about their effect on subsoil hydrology indicates that such a relationship may exist.

Root development increases noncapillary porosity; at the same time noncapillary porosity facilitates root development. Soils with noncapillary porosity of 5 percent or less are almost impermeable to roots. In the forest soil illustrated in figure 12, it may be assumed that tree roots penetrated depths where the volume of noncapillary pores varied between 30 and 12 percent. But below 18 inches, where such pore space was only 3 percent, no roots were found. In the agricultural soil, it may be assumed that the roots were confined to the upper 6 inches because noncapillary porosity below this depth was 3 percent.

In compacted soils, the reduction in noncapillary porosity hinders root development. Tests indicate that sunflower roots do not penetrate soils that have been compacted to volume weights ranging from 1.75 for sands to 1.46–1.63 for clays (*38*). The effect of a dense soil on root development is shown in figure 15. The B horizon of the Butler silt loam in which the western wheatgrass was growing was so compact that root growth was greatly inhibited. The C horizon, however, was less dense, allowing normal root growth.

Lutz (*31*) has cited research in Europe and the United States indicating that trampling in woodland parks compacts the soil and kills roots. Another study (*26a*) has shown that woodland grazing reduced tree growth as much as 50 percent, probably attributable in large part to decreased root development caused by compaction.

These investigations all confirm that compaction not only may reduce infiltration and percolation capacities by decreasing noncapillary porosity, but may also reduce percolation capacity by hindering root development.

Since the volume of roots has received relatively little investigation, it is difficult to estimate their effect on noncapillary porosity. Although the ratio of amount of material produced in

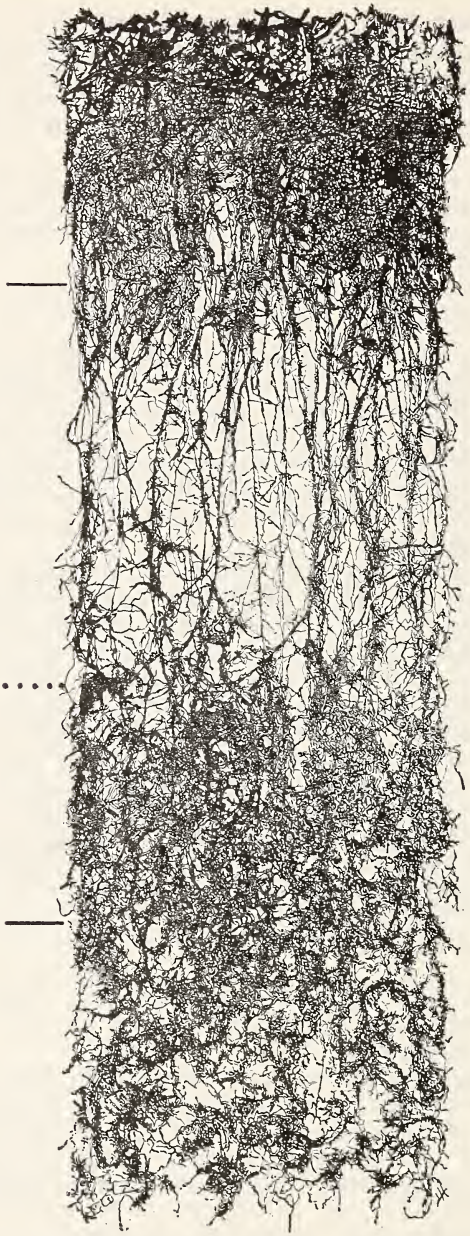

FIGURE 15.—Root system of western wheatgrass from a claypan soil, Butler silt loam, showing great differences of root habit in the three soil horizons. There are fewer roots in the shallow A horizon (above upper line) than is normal in mellower soils. Branching in the B horizon (above lower line) is poor except in the lower third. Branching is pronounced in the mellow, moist soil of the C horizon. Dots indicate a depth of 20.5 inches (*41*).

vegetation above ground to that below ground varies considerably, the indications are that roots occupy a considerable volume. Kittredge (*28*) has utilized data from a study by Duncan (*12*) to calculate top-root ratios for one-year tree seedlings on the same soil: red gum, 2.31; eastern redcedar, 0.92; northern red oak, 0.45; and post oak, 0.24. These data show that roughly 0.5 to 4 times the amount of material growing above ground is growing below the surface. These values are on a weight basis; on a volume basis the proportion of material below ground may be even greater, since much of the volume is composed of root hairs which are lighter in weight than the more woody material.

Dittmer (*10*) has reported that the hairs on the roots of rye quadruple the volumes of soil from which the roots absorb moisture. An indication of the volume of roots is also provided by the length of roots in a cubic inch of soil: Oats, 3.5 feet; rye, 5.0 feet; bluegrass, 30 feet, with external surfaces of 15, 30, and 65 square inches, respectively. Figure 9 also indicates the large volume of soil occupied by roots.

Gaiser (*17a*) has recently found that taproot cavities in an oak-hickory stand in southeastern Ohio penetrate as much as 6 feet and number an estimated 4,000 per acre. Kramer (*29*), reviewing the literature on root concentration, cites many other references indicating the large volumes of roots. This volume, in large part, may be considered as an important source of noncapillary pore space. The root-soil interface, when occupied by the living root, is an avenue of gravitational moisture movement. Death and decay of roots and subsequent filling in of the cavity with surface material gives the root cavity a greater noncapillary porosity than the surrounding soil.

Relation of Percolation to Subsurface and Base Flow

Many observations confirm that when the percolation capacity at some layer is less than the infiltration rate, lateral downslope flow will occur in the soil above that layer. In one test, for example, an infiltrometer run was made on a 20-percent slope on a soil with a compact layer at 28 inches depth. A well to the compact layer was dug 15 feet downslope from the infiltrometer. Rainfall was applied to the infiltration plot at a rate of 3.5 inches per hour for a period of 1⅔ hours. Infiltration during this period averaged 3.2 inches per hour. Fifty minutes after the artificial rainfall began, enough water had seeped into the well to fill it to within 10 inches of the surface. At the end of rainfall application the well-water surface was within 7 inches of the soil surface.[7]

The rate and quantity of lateral or subsurface flow is determined largely by the characteristics of the noncapillary porosity of the soil and by the hydraulic gradient of the flowing water. In some soils the rate of subsurface flow is so slow that for all practical purposes it contributes nothing to stream flow during

[7] Original data used in preparing Station Paper No. 39, EFFECT OF SOIL AND COVER CONDITIONS ON SOIL-WATER RELATIONSHIPS. 1951. U. S. Forest Serv. Northeast. Forest Expt. Sta.

periods of flood runoff, and is therefore not reflected in the storm hydrograph. In other cases subsurface runoff is quite rapid, and an appreciable amount may reach the stream before all channel flow originating from surface runoff has moved downstream.

As mentioned earlier, the movement of soil water toward the channel is slower than surface runoff. There are several reasons for this. Where surface runoff occurs the soil surface is often honeycombed with small channels, rills, and depressions directly tributary to the larger channels that carry water continuously during rainstorms. It is very seldom, therefore, that surface water has to travel far—relatively speaking—before it reaches some watercourse. Below the surface, however, water moving laterally as subsurface flow, or vertically towards the permanent water table, takes longer to get to the stream. Resistance to flow along a unit length of pores is usually greater than that of the surface, because contact with more particles and with a greater soil area develops greater friction.

The deeper the water penetrates, the slower is its movement to channels. All flow below the levels of rill and gully incisions must continue in the soil until an impermeable layer is reached, and then it travels a longer distance before reaching the stream channel. In gullied areas, lateral flow may be quickly intercepted by a gully, and then its transit to the stream is speeded up.

Temporary (detention) storage in the noncapillary pores also provides an opportunity for deep seepage. As indicated by figure 14, for example, deep seepage below the 30-inch depth is possible in the soil represented by curve *1* at the percolation rate of 0.02 inch per hour. A storage period of 24 hours will permit 0.48 inch to flow to the underground reservoir. A storage period of 48 hours, not uncommon, will permit 0.96 inch to go to deep seepage. This is the water that moves most slowly. Once in the underground reservoir, it may take months, even years for the water finally to reach the stream.

APPLICATION TO WATERSHED MANAGEMENT

Every unit of land, however small, is part of a watershed. Hence the manner in which the land is managed will inevitably have some influence on stream flow, as the previous discussion of plant-soil-water relations has indicated.

These relations provide a rational means of evaluating the effects of given forest and range management practices on stream flow. By properly interpreting soil and plant cover conditions, the land manager may predict with greater assurance the effects of thinnings, harvest cuttings, logging, grazing, revegetation, prescribed burning, and other treatments. Practical application of the knowledge concerning these relations may also open up some interesting possibilities for special practices to achieve water-control objectives. Thus the knowledge that root depths affect quantities of transpiration and storage may suggest effective methods of controlling water yield.

Scientific land management in its broadest sense has not been

possible so long as the influence of various practices on stream flow—expressed through plant-soil-water relations—has not been adequately understood. These relations provide an important basis for determining forest and range management objectives and practices throughout the Nation. This is true because the basic principles of movement and storage of soil water apply everywhere. Vegetation, wherever it grows, removes water from the soil, and through its roots and organic matter affects the movement and storage of water. Stream flow in turn, wherever it may occur, reflects the interaction of soil and vegetation.

Basic Information Needed

Application of the principles of watershed management requires specific information as to the conditions affecting plant-soil-water relations for the watersheds to be managed. First of all, broad regional differences, such as in climate, must be considered. Thus, frost in the Northeast, heavy snow packs in the western mountains, and low precipitation in the Southwest represent regional problems to be dealt with within the framework of the same basic principles. Secondly, consideration must be given to the local variations in climate, vegetation, and soil characteristics (including significant geologic factors, among others) of the specific watersheds to be managed. It must be reemphasized, however, that irrespective of the regional and local variations the fundamental relations that have been described apply in all instances.

Most forest and range watersheds are needed for timber and forage production. However, as our needs for the conservation and control of water have assumed greater urgency it has also become necessary to include flood control, water production, or both, among the major management objectives. For these reasons a minimum amount of hydrologic data will be required for most—if not all—managed forest and range areas.

Basic data may be conveniently placed in three major categories: (1) Water, (2) soils, and (3) vegetation. Data required under item 1 include the precipitation over the watershed, and the proportion represented by stream flow. Knowledge of precipitation characteristics should include types and intensities of rainfall, and the amounts that fall as rain or snow. Data should be collected on the distribution of stream flow throughout the year, the uses to which the water is put within and below the watershed, and the problems, if any, associated with stream-flow variation and quality.

In category 2, information should be acquired on the storage characteristics of the several important soil types, especially water-holding capacity, and depth in relation to root depth. The collection of data on the seasonal movement of soil moisture will provide an essential guide to the volume of available storage under different types of vegetation and of land treatment. Infiltration and percolation values will be needed to determine the proportions of precipitation that go to surface, subsurface, and base flow, respectively, as well as the utilization of storage space by soil-water movement. Rates of accumulation and incorporation of organic

matter into the soil will require systematic observation. The preparation of maps indicating the pertinent hydrologic characteristics of the soils will provide necessary "working tools" for the conduct and evaluation of management practices on the watersheds concerned.

In category 3, information will be needed on root depths of plants of various species and ages. A working knowledge of ecology, particularly in reference to the water-using characteristics of various stages of plant succession, is also required as a basis for determining the possibilities of influencing water yield by manipulating the plant cover. In addition, phenological observations will help in providing useful indications of seasonal water use by plants.

The information just outlined is basic to effective watershed planning because it will permit an evaluation of the effects of current forest and range management practices on stream flow. It can also be used to help determine which practices will best meet the requirements for stable conditions of soil and stream flow, and how intensively forest or range areas can be used without endangering these conditions.

It cannot be emphasized too strongly that the data on water, soils, and vegetation must be integrated before they can be applied effectively. Water, soils, or plant data separately are of little practical value. For example, data on soil depth are usable only where interpreted in relation to root depth, and to total soil-water storage and opportunity for storage. Such an integration next brings to light the extent to which storage is utilized by soil-water movement, and finally, the consequent effect on stream flow.

The amount of basic information required for scientific watershed management may appear formidable in relation to the time, training, and facilities currently available to land managers. However, it should be remembered that scientific watershed management, today virtually nonexistent, cannot be learned or applied immediately. For some areas it may take 10 years to obtain the needed information. Meantime the land manager can do much to encourage the development of watershed conditions which will contribute directly to scientific watershed management. For example, he can so plan and conduct forest and range management practices as to maintain the maximum possible stability of the soil. In short, watershed management, as conceived in this publication, involves a change in the land manager's viewpoint. Once he extends his thinking to include soil and water among the basic resources for which he is responsible, watershed management may be said to have begun.

In practice, the types and intensities of watershed practices to be applied to a given forest or range unit will depend upon local conditions and problems. Consequently, the following consideration of some specific applications of fundamental plant-soil-water relations can be only in rather general terms. Flood control will be considered first, water yield second, forest and range management practices third, and finally an example will be given of the practical application of watershed-management measures.

Flood Control

Floods may be divided roughly into two types: (1) The flash flood of high peak and short duration, usually resulting from a convectional type storm on a relatively small area, and (2) the sustained flood of large volumes and long duration caused by the prolonged rainfall of a convergent type storm (often with snowmelt added) which more than satisfies the retention storage capacity of vegetation and soils over large areas. Each type of flood can be reduced by watershed management.

The flash flood as a general rule is caused primarily by surface runoff. Hence the first objective is to get more of the water into the soil, and thus into such storage as may be available, or into the slow-moving subsurface and base flows. Because such floods usually occur during the growing season, both retention and detention storage are usually available. Measures designed to increase both types of storage can thereby result in a considerable degree of control over these floods. The fact that the excess water usually comes from a small area adds to the effectiveness of such control. Along the Wasatch Front in Utah, for example, floods of this nature, in six different watersheds, originated on 1,315 acres or only 9.2 percent of the total area of 12,868 acres (*4*).

Since the sustained large flood results from large volumes of rainfall or snowmelt over a large area, the possibilities of its control by watershed management are much less than in the case of flash floods. Infiltration conditions are in general not as critical as in the case of flash floods. Maximum rainfall intensities are also lower. As with flash floods, the primary objective is to get the water into the soil, but because retention storage is already filled in case of the large flood, the water can go only into detention storage. Under these conditions the principal benefit from watershed-management practices is realized by increasing detention storage, thus achieving a reduction in flood flows.

Since the first objective in planning to reduce floods of either type is to facilitate the entry of precipitation into the soil, infiltration capacity is the first limiting factor. Since infiltration capacity depends on the noncapillary porosity of the surface soil, any forest- or range-management practices that modify the surface pore space will exert a direct influence upon infiltration.

Two questions must be considered in evaluating land-management practices. First, do they bare the soil to rainfall impact? Second, do they compact the soil to an appreciable extent? Soil exposure and soil compaction directly affect the noncapillary porosity of the surface soil, and thus its infiltration capacity. For these reasons, the soil must be protected from such disturbances—and to the maximum possible extent. This applies especially to areas where flood control is a paramount need.

Infiltration is also affected by those subsurface factors which determine the rates of water entry into the soil surface. It is therefore important to know whether the constant rate at which water enters the soil is governed by surface or subsurface conditions. In order to judge what measures are needed to increase rates of water movement it is necessary to know whether the rate

reflects the infiltration *capacity* of the upper ½ inch of soil, or an infiltration *rate* determined by the percolation capacity of some deeper level. Thus, the previously explained distinction between infiltration capacity and infiltration rate has more than an academic importance.

Soil-water storage is a second limiting and highly important factor in flood control. Forest and range managers must therefore aim towards maintaining the maximum storage space for precipitation that otherwise would create or aggravate flood conditions. Although this is not an entirely new concept, it needs to be stressed because heretofore the flood-control value of abundant forest cover and well-maintained range cover has been thought of chiefly in terms of maximum infiltration capacity and soil stability.

How much precipitation the soil reservoir can hold depends on several factors. One of the most important is soil depth, for this largely determines how much can be accomplished towards increase of storage.

Flood control on shallow soils.—On first thought, the possibilities of increasing retention storage space in shallow soils would appear to be poor. But the fact that this storage is limited on such areas indicates that small increases in storage can achieve a great deal relatively. One of the most obvious methods is to encourage the maximum possible growth of vegetation. Although such growth on shallow soils will not immediately increase the depth in which storage occurs, it will have two beneficial effects. First, a cover whose roots fully occupy the soil maintains retention storage opportunity at a maximum, thus reducing the volume of water which can go into flash-flood flows. Second, the vegetation protects the soil against rainfall impact and maintains relatively high infiltration and percolation capacities.

These relationships provide further evidence that the encouragement of maximum vegetal growth should be a key objective of management on shallow-soil areas that are sources of floods.

Some deep soil mantles on flood-source areas have shallow subsurface layers impervious to the roots of existing vegetation. These soils operate for all practical purposes as shallow soils. However, it is possible under some circumstances to increase storage by encouraging vegetation that can penetrate compact subsoils. Although little is known about this possibility in the United States, Russell (*35*) has pointed out that roots of some plants can penetrate even quite compact materials. Among other examples he cites the use of lupine in Germany to penetrate hardpans underlying sandy soil, and alfalfa and sweetclover in irrigated areas to improve drainage.

The organic matter added to a shallow soil by a good growth of vegetation increases the total storage space. In such soils the roots of almost all plants will penetrate to the impermeable layer. Hence storage is limited to the zone between the soil surface and the impermeable layer. Examples are the Volusia soil in New York State with an impermeable layer 18 to 24 inches below the surface, and the Toledo and associated soils in the Lake States.

Even a small increase of storage capacity by the addition and incorporation of organic matter may significantly affect runoff from such soils.

This effect is achieved in large measure by the action of organic matter in facilitating aggregation and thus increasing detention storage. The maintenance of a maximum plant cover offers an outstanding possibility in this respect. Additional detention storage is especially valuable during rainy seasons when retention storage space is filled, leaving only the detention storage available for precipitation that might cause floods.

It is probable that most forest and range areas do not now have the maximum amount of organic matter they can acquire because fire, overcutting, heavy grazing, and other conditions have depleted the cover. It seems reasonable to suppose, therefore, that the organic content of the soil may be raised and its depth correspondingly increased.

Billings (*6*) has shown that an increase in organic-matter content of the soil during the growth of old-field shortleaf pine stands was associated with an increase in storage capacity of the upper 2 inches of soil. Prior to establishment of the stand, water-holding capacity was 33.7 percent by volume. By the time the stand was 21 years old the capacity had increased to 42.4 percent. Thereafter the capacity remained fairly constant, reaching a value of 45.3 percent at 110 years.

Increases in organic-matter content will not continue indefinitely, however. As Nikiforoff (*32*) has pointed out, the organic-matter content of a soil builds up to a point where the rate of production of organic matter is balanced by its rate of decomposition. Thereafter, no further accumulation takes place and an equilibrium is established. It is doubtful if such an equilibrium is ever approached in most forest areas, or if so, maintained for any period of time, because fire and cutting would tend to keep the organic-matter content below equilibrium level.

Soil storage capacity may also be affected by changing the type of forest cover. Studies at the Harvard Forest (*18*) indicate how this occurs during successive stages of vegetation from abandoned field to white pine to hardwoods. As the old-field white pine stand developed, a mor formed.[8] This attained a maximum depth of 2 inches by 60 years. After the pine was clear-cut (at 80 years), hardwoods invaded the area. During the first 10 years, the mor layer decomposed to less than an inch while organic matter was incorporated into the mineral matter (as a mull) to a depth of 4 inches. In time, the mull organic layer came to average about 6 inches, and thereafter remained about the same thickness, although the depth of incorporation varied with the species.

Soil penetration tests during the pine stage showed that as

[8] Type of forest humus layer: *Mor*, unincorporated organic material, usually matted or compacted or both, distinctly delimited from the mineral soil, unless the latter has been blackened by washing in organic matter. *Mull*, organic and mineral matter so mixed that the transition to the underlying layer is not sharp.

the stand became older, the soil became more compact. The soil structure changed very little over time, remaining either granular or crumby, with some tendency towards columnar structure as it aged. But as the hardwoods came in the soil loosened up and became friable. The granular and columnar structures altered to a crumby structure. According to Fenton (*14*) a shift toward a mull condition may be obtained by frequent thinning and leaving the green slash on the ground. This results in the formation of a more desirable form of humus and a consequent increase in storage capacity.

Flood control on deep soils.—The storage space of deep soils may be increased in the same manner as that of shallow soils. But there are additional opportunities for increasing their effectiveness in flood control, namely, through encouraging the growth of deep-rooted vegetation. As previously stated, soil moisture is removed to the depth of the rooting system, so that if shallow-rooted plants are replaced by deep-rooted plants, storage space will increase.

Practical possibilities of applying this principle have yet to be studied. Two limitations must be considered. First, the soil must be of sufficient depth and possess such physical properties as will permit the establishment or encouragement of plants with roots deeper than those of the existing cover. Second, the continued maintenance of one type of cover replacing another may require considerable effort to prevent the return of the former vegetation or invasion by other vegetation better adapted to the site.

Insofar as flood control is concerned, maintenance of a well-distributed root system appears to be as important on deep soils as on shallow soils. In this way the opportunity for retention storage of precipitation is kept at a maximum.

The objective of maintaining storage space applies not only to well-stocked mature stands, but also to other stages of stand development. Thus in planting or reseeding flood-source areas, effort should be made to use species capable of developing the deepest possible root systems. The effects of cultural practices, thinnings, and harvest cuttings on such areas should also be evaluated from this standpoint.

The fact that forest cover occupies an area does not necessarily indicate that a favorable water-storage opportunity exists. Condition of the cover as well as that of the soil beneath must be considered. To illustrate, the soil in a grazed deciduous forest on the Pee Dee River watershed in North Carolina was compact, contained little organic matter, and had a low storage capacity.[9] By contrast, a humus layer 8 inches deep had developed on a similar area protected from grazing for 24 years. Although both areas had shallow soils and yielded flood runoff frequently because their limited storage capacity was exceeded, the 8-inch humus layer of the protected area represented an additional storage capacity of 2 inches.

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[9] LASSEN, LEON. FIRE MEMO. (Pee Dee Flood Control Report.) 1949. U. S. Forest Serv., Div. of Forest Influences.

Opportunities for Increasing Water Yield

The greatest opportunity known to date for increasing the yield of water lies in minimizing transpiration losses, e.g., by favoring vegetation with relatively low requirements for water. As previously described, the amount of water transpired by vegetation depends first, upon the amount available, and second, upon the depth of soil occupied by roots. Since moisture availability is fairly constant for any watershed in question, the amount of water transpired will depend primarily on root depth. Shallow-rooted species obviously use less water than deep-rooted species that can tap a greater supply of water. As less water is transpired, more would be available to stream flow. Thus, on deep soils, substitution of shallow-rooted for deep-rooted vegetation offers one possibility for increasing water yield.

A recent 3-year experiment by Croft (*9*) in the Wasatch Mountains in Utah indicated that by the end of each growing season, evapo-transpiration of an aspen stand with herbaceous understory drew moisture from the soil to the extent that 11 inches of retention storage space had to be replenished before water could become available to stream flow. The roots of the aspen extended at least 6 feet deep and those of the herbaceous cover 3 to 4 feet. Under the herbaceous cover alone, storage was provided for 8 inches of water; whereas in a bare area, evaporation had provided storage for only 3 inches of water. Therefore, removal of the aspen alone saved 3 inches of water for stream flow, whereas the removal of all vegetation, including the aspen, saved 8 inches of water.

The same study also brought out the importance of retaining enough vegetation to protect the soil. The dense herbaceous cover on the experimental plots effectively prevented erosion. In sharp contrast, excessive soil losses occurred where all cover was removed. Thus, although the highest water yield was obtained by complete removal of the vegetation, the cost in soil loss was prohibitive. Those who propose denudation to increase water supplies should therefore not overlook the destructive effects upon water values by the addition of silt, as well as on the productivity of the land itself.

Water yield from some forested watersheds may be increased by clear-cutting the vegetation and leaving it on the area so that the slash and debris protect the soil against erosion. Such treatment in the well-known experiment at the Coweeta Hydrologic Laboratory in North Carolina (*22*), increased total runoff 17.29 area inches (or 65 percent) the first year after cutting. Increased transpiration by the ensuing sprout growth reduced this gain to 13.26 inches by the second year. Subsequent annual cutting of sprouts over the past 7 years has maintained the increase in runoff which resulted the first year after cutting.

The experiment in North Carolina further illustrates the results of substituting a shallow-rooted for a deep-rooted cover. Following removal of the trees, shallower-rooted herbaceous ground cover and sprouts took over the site; less water was removed from the soil, and more went to stream flow.

Taking advantage of the close relation between root depth and water yield to increase water supplies is possible only in the case of deep soils fully occupied by live roots. The increment derived by favoring shallow-rooted plants would be equivalent to the resulting change in storage between the field capacity and wilting point levels throughout the affected depths.

Shallow soils, on the other hand, do not permit changes in water yield by means of cultural treatments. For example, a study in New York State of the effects of planting abandoned fields to trees in three small watersheds showed that water yield had not been affected (*3*). The areas had previously supported herbs and shrubs whose roots extended 18 to 24 inches to a hardpan layer. Since the roots of the planted trees could not penetrate the hardpan, no greater volume of soil could be occupied. Consequently, water yield was unaffected.

Partial removal of the cover to increase water yield would tend to increase evaporation because the site would be more exposed to sun and wind. This practice would be effective, therefore, only where the decrease in transpiration was greater than the concurrent increase in evaporation. If the ground cover were disturbed as little as possible during removal of the overstory, the increase in evaporation would probably not equal the decrease in transpiration. Vegetation and litter sufficient to maintain soil stability would also effectively prevent marked increases in evaporation. With shallow soils such increases in evaporation as might occur would be relatively more significant because only a relatively small amount of water can be diverted to stream flow by reducing transpiration losses.

The fact that manipulation of the cover on shallow soils cannot be as effective as on deep soils does not eliminate the former from consideration in seeking increased water supplies. In the higher western mountains, where precipitation comes chiefly as snow, heavy cutting in stands of conifers, such as lodgepole pine, will materially reduce interception and thus make additional water available for stream flow (*42*).

Cutting of riparian vegetation represents still another possibility for increasing water yield that deserves more attention. Since the amount of water transpired depends primarily on the amount available, the heaviest transpiration losses will occur where roots are in continuous contact with a water supply. It follows that water yield may be increased most effectively in such areas. Experiments at Coweeta showed that cutting all riparian woody vegetation in a strip extending to a vertical distance of 15 feet above the stream channel, affecting 12 percent of the total watershed area, increased summer base flow by approximately 12 percent as measured during a 10-day period in June (*13*).

In conclusion, the practical possibilities and limitations of achieving increased water yield must always be considered. First, the manipulation of cover, either by substituting one type for another, or by repeated cuttings to reduce transpiration losses, will usually be laborious and expensive. Such practices are feasible primarily in areas where water shortages are acute and water

values are high. For example, in trying to maintain a shallow-rooted herbaceous cover that would provide soil stability and yet use little water, the problem of arresting the ecological trend towards deeper-rooted cover may have to be faced. Similarly, the favorable growing conditions near stream channels would make the maintenance of increased water supplies an expensive process, aside from the necessity of preventing stream-bank erosion after destroying the cover.

Forest and Range Management Practices

The fact that forest- and range-management practices affect plant-soil-water relations makes it necessary to give more attention to this aspect in timber production and range management than has previously been given. It has been shown, for example, that removal of the cover, whether by cutting or grazing, and the choice of species for seeding or planting will affect soil-water storage and movement. Fire also takes on additional importance when its hydrologic aspects are taken into account. This indicates the desirability of including the effects of fire upon soil-water relations among the criteria for determining adequate protection requirements for a given area.

The influence of these factors will be summarized later. Consideration is given first to an effect accompanying forest and range use which is not due to cover changes alone but to an associated physical effect, soil compaction.

Soil compaction.—Soil compaction may be defined as the packing together of the soil particles by a force exerted on the soil. This force is represented by any weight—that of a person, truck, or animal, or the impact of raindrops on bare soil. Thus, any use of forest or range areas involves some degree of soil compaction. Its principal effect, so far as soil-water relations are concerned, is reduction of pore space corresponding to an increase in soil density. Reduction of pore space is achieved chiefly at the expense of the noncapillary pores—those governing infiltration capacity.

The effects of compaction are confined to the upper few inches. The force varies according to the type of load. On the basis of static loads, sheep exert a ground pressure of about 30 pounds per square inch, cattle 50 pounds. Adult human beings of average weight exert a pressure of 6 to 15 pounds, depending on the type of shoes worn.[10] Crawler tractors exert pressures of from 3 to 8 pounds per square inch, wheel tractors 20 to 28 pounds. Truck pressures range generally from 36 pounds for pickups to 85 pounds for 5-ton vehicles. Giant trailers carrying maximum permissible loads of logs produce pressures ranging from 85 to 95 pounds per square inch.

As might be expected, the area compacted by forest or range use varies considerably. Grazing affects most of the area involved, the degree of compaction depending on the intensity of use. Alderfer and Robinson (*1*) found in tests on a variety of

[10] Lull, Howard W. soil compaction. 25 pp., illus. 1950. [Unpublished manuscript. U. S. Forest Serv., Div. of Forest Influences.]

pasture sites on clay loams and sandy loams, that volume weights of the soil layer 0 to 1 inch deep varied from 1.54 to 1.91 on heavily grazed sites, compared with 1.09 to 1.51 on ungrazed and lightly grazed sites. Noncapillary porosity for the two sites varied from 3 to 10 percent, and 15 to 33 percent, respectively. Compaction was largely confined to the upper inch of soil.

Logging operations compact a smaller proportion of the affected areas than does livestock grazing. Skid roads usually cover only about 10 percent of an area, but the actual area compacted, including loading sites and skid trails, may amount to 40 percent of the total. Little is known to date about the effects of logging on soil compaction, soil-water relations, or stream flow. Its obvious effect is to reduce infiltration rates through increase of soil density. Free, Lamb, and Carleton (*17*) found that one trip with an empty farm truck or tractor increased the density of the upper 2 inches of a silt loam from a volume weight of 1.32 to 1.50. Four trips with an empty truck increased the volume weight from 1.24 to 1.58 (representing an approximate reduction in total pore space from 53 percent to 40 percent). Analysis of the effect of a similar force on the permeability of a soil sample revealed that permeability—as measured by the time it took 0.4 inch head of water to disappear—varied from 1 minute to 192 hours depending on the degree of compaction achieved. The degree of compaction, in turn, was related to soil moisture and organic-matter content.

Compaction of some part is so closely associated with productive use of forest and range areas that little can be done to reduce its effect on soil-water relations. However, its influence on these relations and therefore on stream flow requires that more information be sought on the process, and on methods of ameliorating its more harmful effects. One interesting aspect of soil compaction relates to past research findings on the close correlation of cover density with erosion rates. It is quite possible that the significant increases in surface runoff associated with reduction in plant density may have been due as much to the soil compaction that occurred during the operation of plant removal as to the decrease in plant density itself.

Effects of timber cutting.—The effect of timber cutting on stream flow will depend primarily upon the proportion of the total number of trees cut and the depth of the soil. Commercial clear-cutting would increase water yield in proportion as root depth is affected and the soil surface is disturbed and compacted by logging. Even in clear-cutting, however, it would rarely be possible to increase water yield by an amount equivalent to the storage capacity of the depth of the soil formerly occupied by live roots.

The actual increase will be influenced, of course, by the root depths of the felled timber; that is, the shallower their roots, the less the effect on water yield. With removal of most of the timber, evaporation would increase as would also the use of water by the remaining ground vegetation. Both of these would tend to reduce the effect of the cutting on water yield.

Soil compaction associated with logging will increase water yield to the extent that surface runoff is increased at the expense of soil-water storage. Such an increase cannot, in general, be regarded as beneficial however, since it is usually accompanied by sediment from accelerated erosion.

If a clear-cut area is allowed to return to forest cover, as it usually will unless heavily or repeatedly burned or converted to cropland, a steady reduction from the initial gain in water yield may be expected. Deeper soil penetration by roots of new growth will gradually increase the retention storage space and consequently reduce the amount of water available for stream flow. As the vegetation thickens, insolation and wind movement will be reduced, thereby lessening evaporation, but this will be more than offset by greater transpiration. Finally, as the roots regain the same degree of development as that of the former stand, stream flow will again shrink to its previous yearly volume. This process may require a number of years—perhaps 30 years, according to one study.[11]

It is also possible that a clear-cut area may be invaded by vegetation with roots deeper than those of the previous growth. In such an event, the clear-cutting would result in an eventual reduction in water yield.

Serious soil disturbances produced by logging or other factors may considerably lengthen the period over which cutting affects water yield. In such cases, restoration of previous water-yield conditions would not be expected until the area was completely revegetated and soil porosity restored. In areas where severe erosion follows denudation, soil-water storage may be greatly reduced both by a decrease in soil volume—hence in total storage space—and by a draining out of water from underground storage by gullies.

The influences of less drastic types of cuttings—for example, thinnings and partial harvests—will depend on the intensity of the cut. Heavy thinnings will reduce the drain on soil water more than light thinnings, but not proportionately. Twice as great an increase in yield cannot be expected from a thinning which removes twice as many stems as another thinning. Light thinnings can be expected to have little or no effect on water yield, since the remaining trees will soon be able to utilize the water previously extracted by those removed. To be effective a thinning must result in the death of roots within masses of soil which do not also contain the live roots of the remaining vegetation. Eventually new root growth will penetrate these soil masses, but until then, water yields will remain higher than before cutting.

Partial harvests such as shelterwood or selective cuttings will tend to have effects similar to those resulting from thinnings. One difference to consider, however, is that thinnings, by definition, apply to young stands which are usually much more capable of accelerated stem and root growth following their release than are older stands. Not only is more moisture available for the

[11] Unpublished data, Coweeta Hydrologic Laboratory. 1950. U. S. Forest Serv., Southeast. Forest Expt. Sta.

remaining trees because of reduced root competition, but any soil masses which may be unoccupied by living roots are penetrated within a short period. Partial harvest cuttings, on the other hand, usually involve older trees. Except for seedlings and saplings, the residual stand may not respond so rapidly to the reduced competition; so that any increases in water yield resulting from cutting may persist for a longer time.

Thinnings and partial harvest cuttings both reduce the interception of rain or snow. This is a particularly important factor in regions where heavy snowfall occurs, as Wilm and Dunford have shown (*42*).

Along streams, where competition for moisture is not so severe, increases in water yield will vary much more directly with the amount of vegetation killed or defoliated than in upland areas. Since such vegetation has access to more water than it needs for growth, and therefore transpires at maximum rates, its destruction may tend to reduce the transpiration draft in direct proportion to the reduction in cover (leaf area).

Revegetation.—The effect of revegetation on water yield will depend on the cover already present, the soil depth, and the root habits of the planted or seeded species. If the area has no protective cover and the proportion of surface runoff is high, successful revegetation will reduce water yield. Infiltration will be increased, less water will go into surface runoff, and more will enter the soil. Part of the water entering the soil will be stored for later evaporation or transpiration. These processes will decrease the amount formerly available for stream flow. The reduction in water yield can be considered as the cost of achieving more stable stream flow, higher-quality water, reduced sedimentation rates, and stabilized soils.

Most sites selected for seeding or planting already have some type of cover. The effect of revegetation depends on the difference between the amount of water being consumed by the existing growth and that which will be consumed by the planted or seeded cover. Here again, soil and root depth must be considered. If the soil is shallow and already fully occupied by existing growth, revegetation will not measurably affect water yield (*3*). Under such conditions, all of the available water would be used either with or without new vegetal cover. The new plants could not draw on additional supplies; on the contrary, they would compete with the existing growth for the same supply.

If on the other hand, the soil is deep and the new plants are deeper-rooted than the existing species, we may expect a reduction in water yield. This would not occur until the roots had penetrated and fully occupied the soil at a depth below that of the original cover.

Fire.—The effect of fire will vary with the quantity of vegetation destroyed. This, in turn, will depend on the intensity and type of fire. Here too, soil depth, the rate of establishment of new growth, and its root characteristics will determine the period and magnitude of changes in water yield. An intense fire on a shallow soil previously subject to evaporation throughout its

entire depth would have comparatively little effect on water yield, even if all vegetation were destroyed. Differences would be restricted to the effects on surface runoff and interception. On the other hand, burning on deep soils occupied by deep-rooted vegetation would affect water yield to a much greater extent, because of the resulting changes in retention storage opportunity as well as in interception and surface runoff. Thus a crown fire, by greatly reducing transpiration, would considerably curtail the extraction of soil moisture. A crown fire will also reduce interception. In areas of heavy snowfall this may cause an appreciable increase in snow storage.

Light ground fires may have no immediate effect if the vegetation is largely undamaged. However, frequently repeated light fires will in time affect soil depth, water-holding content, and storage capacity by preventing the accumulation and incorporation of organic matter. In some cases, prescribed burning (which is rapidly growing in favor as a silvicultural tool) requires repeated light fires. This practice, by reducing the capacity of the soil to retain precipitation, may result in more rapid and higher flood discharges from heavy rainfall. Where prescribed burning calls for only one fire, as for the purpose of killing undesirable vegetation, damage to the soil may be slight. From the standpoint of surface runoff, the effect of such a burn will be temporary in comparison with the lasting effect of changes in depth of root penetration that may occur with changes in type of vegetation. For these reasons the use of fire as a tool in forest and range management requires careful consideration. The root and soil depths in the proposed area should be determined, and the effect on water yield evaluated. To these considerations, of course, must be added the effect of fire on erosion, a subject not considered in detail in this circular. Possibilities of soil loss and reduced water quality due to burning may far outweigh potential gains in water yield.

Grazing.—Grazing affects plant-soil-water relations in a number of significant respects, depending upon the intensity of use and type of cover, among other factors. Whereas the effects of timber cutting depend largely on how rapidly new growth returns, the effects of grazing will depend largely on the degree to which the forage is utilized.

The quantity of herbage left after grazing will govern the growth and development of forage plant roots. The rate of root development depends on the availability of carbohydrates manufactured by the plant and not needed immediately for top growth (*35*). Heavy grazing reduces surplus carbohydrate production. Those carbohydrates which are produced are needed primarily to replace the parts that have been cropped. Root growth—and retention storage availability—are reduced accordingly. Measurements on semiarid range land in Utah show that roots of bunch wheatgrass on an overgrazed area penetrated to an average maximum depth of 44 cm. in contrast with about 65 cm. on protected areas. Based on weight of roots per unit volume of soil, root development in the soil of an ungrazed area was

found to be six times greater than in the heavily grazed area (*19*). Flory and Trussell in a study of forage grasses found that the roots of black and blue grama in New Mexico penetrated 4 feet under moderate grazing, about 2 feet under heavy grazing, but to less than 1 foot on very heavily used range (*15*).

Grazing also affects infiltration. As a general rule infiltration capacity is directly proportional to the amount of herbaceous material left. It is therefore possible that where soils are heavily grazed, their already low storage capacities may be only partly utilized because water cannot enter the soil rapidly enough.

The close correlation of vegetation density with infiltration permits the establishment of protection requirements in cover density values as a means of regulating livestock use. For example, research in high-altitude ranges in Utah and southwestern Idaho indicated that at least 65 percent of the soil surface must be covered by plants or litter in order to control surface runoff and maintain stable soil (*16*). This is the density of cover required to protect the soil from rainfall impact with its damaging effect on noncapillary porosity and infiltration capacity, to produce a root system which will facilitate the entry of water into the soil, and to prevent the formation of continuous rills across the soil surface.

The grazing of deciduous forest areas markedly affects plant-soil-water relations. The soil in deciduous forests usually lacks the protective grass or other herbaceous cover with its mat of fine roots that is characteristic of well-managed range areas, and is therefore more susceptible to compaction. Experiments at the Coweeta Hydrologic Laboratory showed that woodland grazing by cattle reduced total porosity of the 2-4 inch depth as much as 50 percent over an 8-year period during which cattle were present only during the summer months. Permeability of the soil was reduced 91 percent.

Application of Principles

A working knowledge of the essential interrelations of soil, cover, and water on given areas can be employed to good advantage in carrying out desired objectives of watershed management. This may be illustrated by a few hypothetical cases in which the need for larger water supplies calls for practices that will favor an increase in useful water flow.

(1) Main source of precipitation is snow; soil is shallow, light-textured; forest growth consists of dense stands of even-aged commercial conifers. In this case, cutting the timber to create openings having diameters equivalent to the height of the dominant trees would reduce interception markedly, yet leave enough shade to prevent excessive evaporation losses.

(2) Main source of precipitation is snow; soil is deep, medium-textured; cover is dense aspen with herbaceous understory. The indicated practice would be to thin the aspen heavily except on wind-swept slopes, where lighter cutting would be desirable in order to minimize the blowing of snow, induce snow accumulation, and extend the period of snow melt.

(3) Precipitation comes as winter and spring rains; soil is deep, heavy-textured; surface thinly vegetated, actively eroding. This condition would call for revegetation with shallow-rooted grasses, brush, or trees, and use of contour trenches and water-spreading devices where necessary to hold the soil and encourage percolation. However, where slopes are very steep, and subject to large amounts of rain, and where subsurface layers are compact, control of soil instability may become the primary consideration, requiring the use of deep-rooted species at first in order to anchor the soil more firmly and increase its percolation capacity.

(4) Rainfall is well-distributed; soils are deep, light-textured; growth consists of merchantable pine-hardwoods. Here the objective would be to maintain a predominantly hardwood stand, favoring those species with shallower roots. A heavy litter would help reduce evaporation losses.

(5) Commercial timber stand is located in riparian zone on alluvial soil; ample water is available throughout the growing season. Although complete removal of the timber would favor the maximum increase in water yield, other considerations, including protection of stream banks against erosion, as well as fish, wildlife, and recreational values, would suggest less drastic measures. For these reasons it would be preferable to reduce the vegetation heavily only in discontinuous strips and to substitute shallow-rooted species wherever possible. Care in harvesting wood products would be necessary to avoid the danger of stream-bank erosion. Much of the stream side would be kept shaded to minimize excess evaporation from the water surface, and to maintain temperatures suitable to native aquatic life.

The illustrations given above exemplify some of the many combinations of factors which must be taken into account in attaining favorable conditions of water flow from watershed areas. It is worth emphasizing in this connection that the application of watershed management principles to actual situations does not necessarily call for drastic shifts in land use or drastic adjustments in timber, forage, and wildlife management practices. Also, in many areas, present and foreseeable requirements for water supplies can undoubtedly be met by adopting or continuing such basic practices as will suffice to maintain a stable soil and consequently the balanced plant-soil-water environment necessary for the continuous production of forest and range resources. In still other areas, climatic, vegetal, and soil factors may be such as to make futile any attempts to increase the yield of stream flow by manipulation of plant growth.

What is important is that the land manager add to his "stock-in-trade" a systematic knowledge of what water services the different parts of his land unit are capable of providing, how and when they are provided, and to what extent they can be improved or increased. He will then be able to predict more accurately the results he is likely to achieve in each subunit by modifying his current management practices or by applying special treatments to enhance the production of good-quality water.

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